Marine and terrestrial geology and geophysics

Shackleton Glacier Project, 1995–1996

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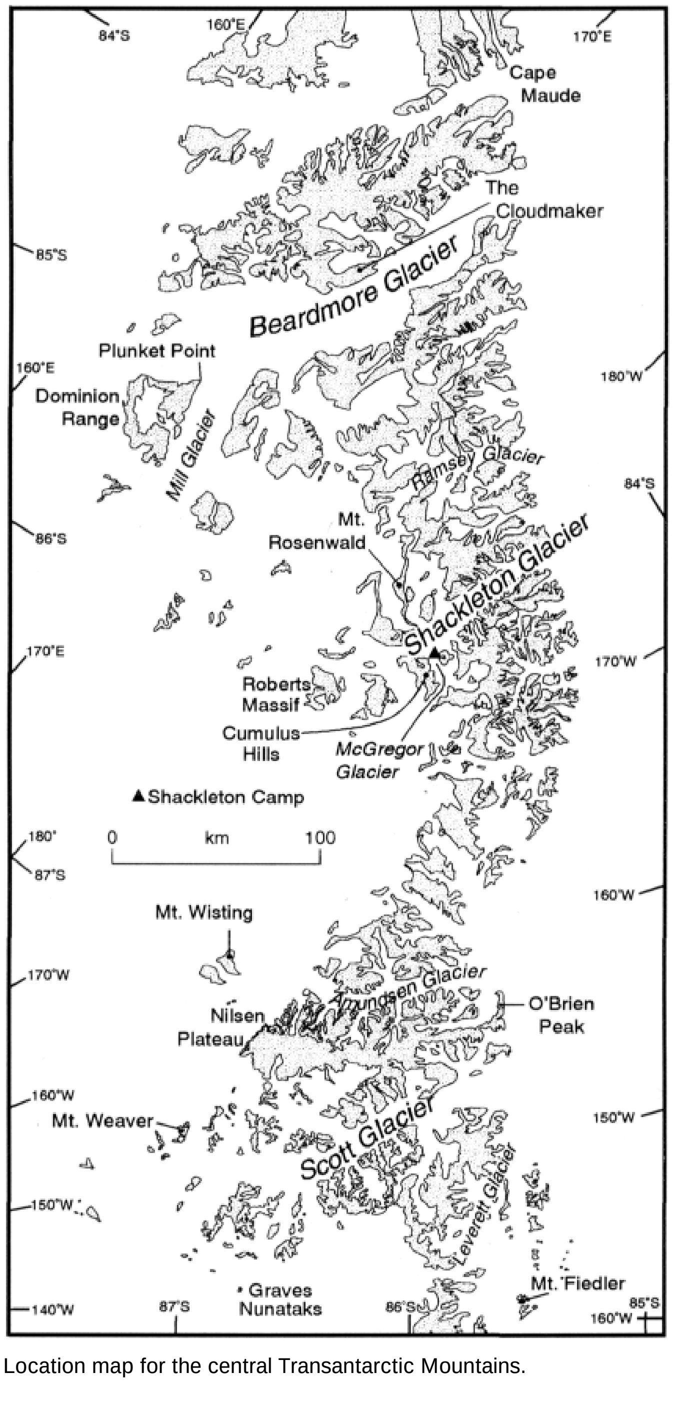
Geological investigations in the Shackleton Glacier region (figure) under the auspices of the U.S. Antarctic Research Program were first conducted by Texas Technological University (Wade et al. 1965; LaPrade 1969). Building on these investigations, a helicopter-supported research program was carried out by geologists from Ohio State University in the 1970–1971 season (Elliot and Coates 1971). Subsequently, surface-supported field projects investigated the vertebrate-bearing beds in the Cumulus Hills (Collinson, Stanley, and Vavra 1978; Cosgriff et al. 1978) and basement rocks along the mountain front (Stump 1975; Borg et al. 1987). Although access can be gained to rocks along the mountain front and near McGregor Glacier, crevassing can make travel dangerous.

The potential for further productive research in the region led a number of investigators to submit proposals for fieldwork to be supported for 2 months by helicopters. Proposals were funded for the 1995–1996 field season, and 8 weeks of helicopter support was approved. The camp structures were brought to the Shackleton Glacier site from the central west antarctic camp in January 1995. On 23 October 1995, a 15-member Antarctic Support Associates (ASA) construction crew arrived at the site. The camp was ready ahead of schedule, in large measure due to prestaging of the camp structures and calm, albeit cold, weather during construction. The helicopters, belonging to Helicopters New Zealand, arrived on 15 November with the first geological groups arriving on 19 November. Helicopter operations terminated on 10 January, and the last science groups returned to McMurdo on 12 January. The camp was then taken apart and transported to Siple Dome in preparation for the 1996–1997 aerogeophysical and ice-core drilling programs.

Science activities

Research was concentrated on the Gondwana sequence, the Cenozoic paleoclimate record, and the tectonic evolution of the Transantarctic Mountains. Science personnel numbered 46, representing 12 different projects. Significant paleontological discoveries were made, including abundant new Lower Triassic vertebrate material from the northern Cumulus Hills and Upper Triassic silicified plant material from near Schroeder Hill. Detailed studies were made on trace fossils in fine-grained Permian beds and on the sedimentology of the coal-bearing Buckley Formation. A system-



Location map for the central Transantarctic Mountains.

atic collection of samples from Permian and Triassic beds was made for palynostratigraphic studies. Detailed investigations were made of paleosols, which are locally abundant, and significant effort was put into locating the Permo-Triassic boundary. Systematic collections were made of the dolerite sills for geochemical analysis and age dating, and the paleovolcanology of the co-magmatic breccias of the Prebble Formation was studied in detail. Continuing studies on the Sirius Group sediments on the Dominion Range resulted in the discovery of peat horizons, and the collection of more material for microfossils and plant remains. These rocks as well as the younger moraines were sampled for exposure age dating. The Sirius deposits were further investigated on Roberts Massif and Bennett Platform. The basement rocks along the mountain front were sampled for paleomagnetic investigations and age dating and for thermochronologic studies of the denudation history. Results of fieldwork by individual projects are reported in the articles that follow.

Field operations

Two Squirrel helicopters, an A model with a payload of about 400 kilograms (kg) and a B model with a payload of about 700 kg (actual payloads depend on altitude and flying distance), operated by Helicopters New Zealand, provided 8 weeks of logistic support for science. In addition, a Twin Otter, owned and operated by Kenn Borek Air, provided 2 weeks of fixed-wing support.

In total, 682 hours of helicopter time were flown in 57 days of operations:

- 576 hours devoted to direct support of earth-science research over a 52-day period;
- 52 hours in support of radar sounding of the Leverett Glacier as part of the investigation of possible overland tractor train routes to the South Pole;
- 10 hours supporting global positioning system (GPS) ground control stations for aerial photography of the Bennett Platform and Roberts Massif and reoccupation of two primary stations established during the 1962–1963 geodetic and topographic survey of the Transantarctic Mountains;
- 26 hours of transit time from McMurdo to Shackleton and return; and
- 18 hours of miscellaneous support—such as installation and recovery of a repeater on top of Mount Rosenwald (3,540 m).

Helicopter operations were under the direction of Ken Tustin, who was supported by three pilots and two mechanics. The 6-hour transit from McMurdo to the Shackleton Camp and return required two refuelling stops: at Senia Point 16 kilometers south of Byrd Glacier and at Cape Maude about 56 kilometers northwest of Beardmore Glacier. Fuel caches were placed on the Mill Glacier by LC-130 aircraft and at various points along the mountain front by helicopter. Only 8 days were lost entirely to weather, including a 4-day break in mid-December and a 2-day shutdown in early January. Flying was curtailed on other days but, except for the ramp-up at the beginning of the season, was never less than 5 hours a day. Excluding nonflying days, 576 hours were flown in 44 days with an average of 13.1 hours per day and a maximum on any 1 day of just over 29 hours. Two sets of crew facilitated science support, and not uncommonly, a day shift was followed by an evening shift. The latter was often used for support in the Dominion Range, which is sufficiently far from the Shackleton Camp (about 170 kilometers to the Mill Glacier fuel depot) that, for safety, the helicopters operated as a pair. Evening operations also supported camp moves and cargo retrograde.

A Twin Otter was based at the camp for 6 days in late November and for an additional 5 days in mid- to late December although that support was somewhat curtailed by poor weather. The Twin Otter enabled visits to places beyond normal helicopter range: Mount Fiedler, Nilsen Plateau, Mount Weaver, O'Brien Peak, and the Dominion Range (figure). In addition, aerial photography of the Roberts Massif and Bennett Platform was flown during the first week and photography of the Dominion Range and The Cloudmaker, during the second period. The Twin Otter also supported the meteorite collection program conducted in the Grosvenor Mountains, near Mount Wisting and near Graves Nunatak (figure).

Camp operations

Camp staff personnel numbered seven: Kevin Killilea (camp manager), two mechanics, one weather observer and radio operator, two cooks, and a camp mountaineer who, as a registered nurse, provided local medical support. The camp mountaineer provided invaluable field support for one of the science projects. The population at Shackleton reflected the number of projects that were operating out of satellite camps and the schedule of arrival and departure from the field. The camp population attained a maximum of 38 (ASA, helicopter crew, Twin Otter crew, and 22 scientists) for a few days in late December, but for most of the time, fewer than 16 scientists were in camp.

Camp facilities consisted of eight Jamesways, including a 12-section science hut, a 16-section galley, a 10-section recreation/radio/washing facilities hut, a 12-section visitors' hut, a 10-section berthing hut for the helicopter crew, and three berthing huts for ASA personnel. All science parties used tents while at the Shackleton Camp. Two prefabricated buildings housed generators and the mechanical workshop. Two helicopter pads were installed for the Squirrels, and a 76,000-liter bladder provided fuel storage. A 2,750-meter skiway was regularly groomed, raising the cargo load of the LC-130 aircraft eventually to about 15,890 kilograms.

Support for D.H. Elliot was provided by National Science Foundation grant OPP 94-20498 to Ohio State University. The success of the camp was made possible by the invaluable and enthusiastic support of the U.S. Navy VXE-6 squadron, Helicopters New Zealand, and ASA personnel.

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Permian to Jurassic palynological collections in the Shackleton Glacier area

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During the period from November 1995 to January 1996, over 700 samples were collected for palynological study from the Shackleton Glacier area. The focus for sample collection and stratigraphic studies is to establish a more precise biostratigraphic framework for the Permian to Jurassic Victoria Group of the Beacon Supergroup in the central Transantarctic Mountains than is currently available. The fossil spores and pollen recovered from these rocks also provide information on the past vegetation and environmental history in this high paleolatitude part of Gondwanaland.

Fieldwork was carried out with helicopter support from a base camp at the junction of the Shackleton and McGregor Glaciers (SHG on figure 1) and from camps at McIntyre Promontory and Graphite Peak. Thirty-eight sections were measured and sampled, and additional samples from other sections and sites were provided by John Isbell, David Elliot, and Greg Retallack.

Samples from the Pagoda, MacKellar, and Fairchild formations (figure 2) were collected from sites on Sullivan Ridge on the Ramsey Glacier, Reid Spur, and Mount Butters (figure 1). The Buckley Formation was sampled at McIntyre Promontory, the upper Ramsey Glacier, Graphite Peak, Mount Finley, and various sites adjacent to the McGregor and Gatlin Glaciers; the uppermost Buckley beds and the Fremouw Formation were sampled at numerous sites at and near Halfmoon Bluff, Collinson Ridge, Shenk Peak, Ellis Bluff, Schroeder Hill, Kitching Ridge, Mount Rosenwald, Layman Peak, Mount Boyd, and Graphite Peak; and the Falla Formation was sampled near Schroeder Hill and at Roberts Massif. Additional samples from the Permian formations were collected from Cape Surprise, McIntyre Promontory, Mount Heekin, and Mount Butters by John Isbell; Permo-Triassic boundary beds from Graphite Peak by Greg Retallack; and the Prebble Formation from Otway Massif and Mount Pratt by David Elliot.

In general, the level of thermal alteration of organic matter is relatively high for the Permian units and somewhat

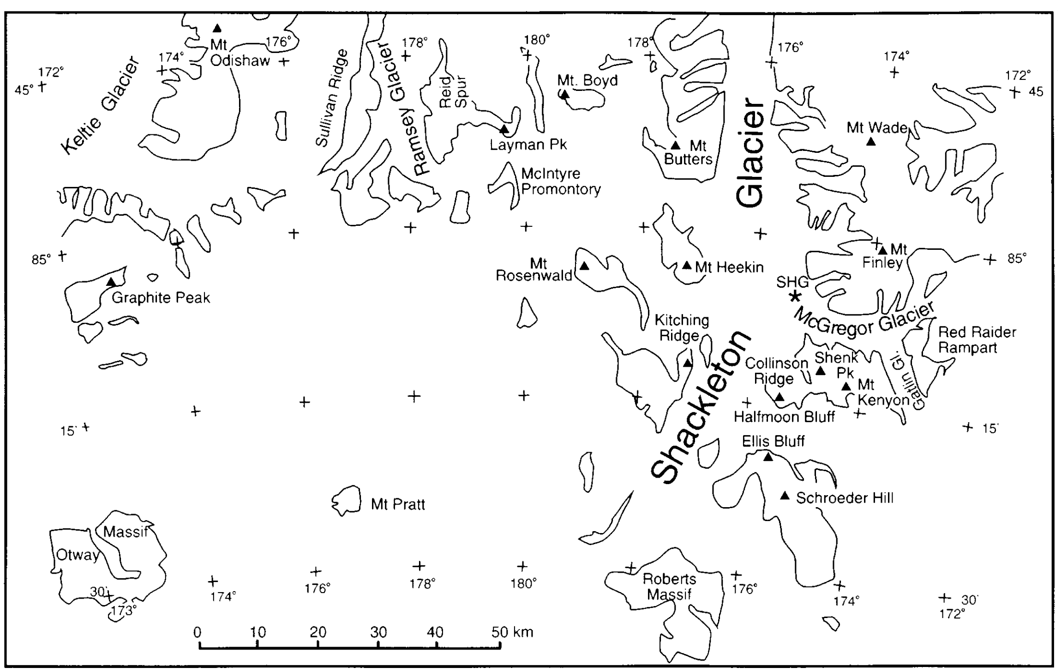

Figure 1. Locality map for the upper Shackleton Glacier area.

higher than experienced in much of the Beardmore area in the Triassic/Jurassic formations. Careful laboratory treatment and different types and amounts of oxidation treatment (as also found by Farabee, Taylor, and Taylor 1991) are helpful in recovering identifiable palynomorphs. Thus far, dark shale samples from the Pagoda Formation contain only traces of organic matter with no recognizable palynomorphs, the dark color probably resulting from iron-titanium oxides. Additional shaly Pagoda samples, however, as well as many samples from the overlying MacKellar and Fairchild formations are still to be processed.

The best-preserved Buckley palynomorphs obtained to date, though black in color, corroded, and barely recognizable, are from the lower part of the formation adjacent to the upper Ramsey Glacier. They appear to be Early Permian and include monosaccate pollen of *Potonieisporites* and *Plicatipollenites*, and taeniate bisaccate pollen, mainly *Protohaploxypinus*.

Triassic assemblages from high in the Fremouw Formation on Layman Peak and from the Fremouw and Falla Formations in the Shroeder Hill area are the best preserved from the Shackleton area thus far. The assemblages from Layman Peak, which contain relatively common lycopsid spores of *Aratrisporites* spp., including *A. parvispinosus* and *A. wollariensis* (figures 3*C* and 3*D*), can be included in subzone B of the *Alisporites* zone (Kyle 1977; Kyle and Schopf 1982) and correlated with the eastern Australian Middle Triassic *Aratrisporites parvispinosus* zone of Helby, Morgan, and Partridge (1987). These results are consistent with previous palynological evidence for correlation and age of the upper Fremouw Formation in the Beardmore Glacier area (e.g., Kyle and Schopf 1982; Farabee, Taylor, and Taylor 1990). The Shroeder Hill samples contain typical Middle-Late Triassic assemblages, with abundant bisaccate pollen (*Alisporites*) and a variety of trilete spores. One specimen of *Polycingulatisporites crenulatus* (figure 3*B*) was found in a sample (AE-12) from an unnamed ridge southeast of Shroeder Hill, suggesting correlation with antarctic subzone D and the eastern Australian *P. crenulatus* zone (de Jersey 1975; Helby et al. 1987), and a Late Triassic age for these beds. The samples from both Layman Peak and Shroeder Hill contain common *Uvaesporites verrucosus* spores (figure 3*A*), and these predominate in sample AE-12. Abundance of these spores, which have a probable lycopsid affinity, and other common lycopsid spores such as *Aratrisporites*, highlight the importance of lycopsid

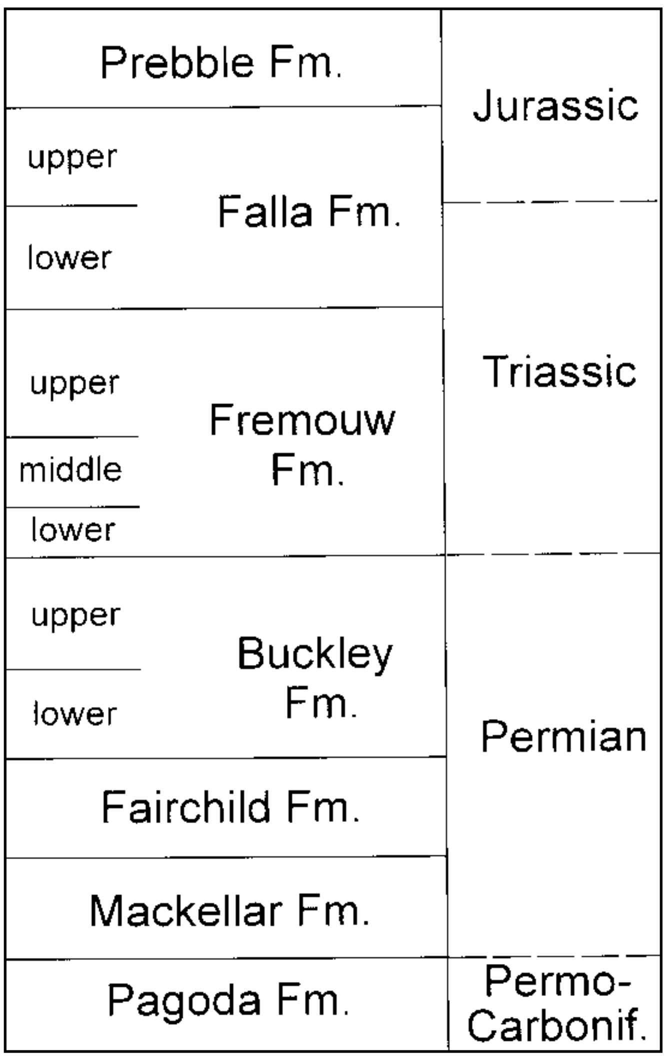

Figure 2. Stratigraphic column for the Victoria Group of the central Transantarctic Mountains.

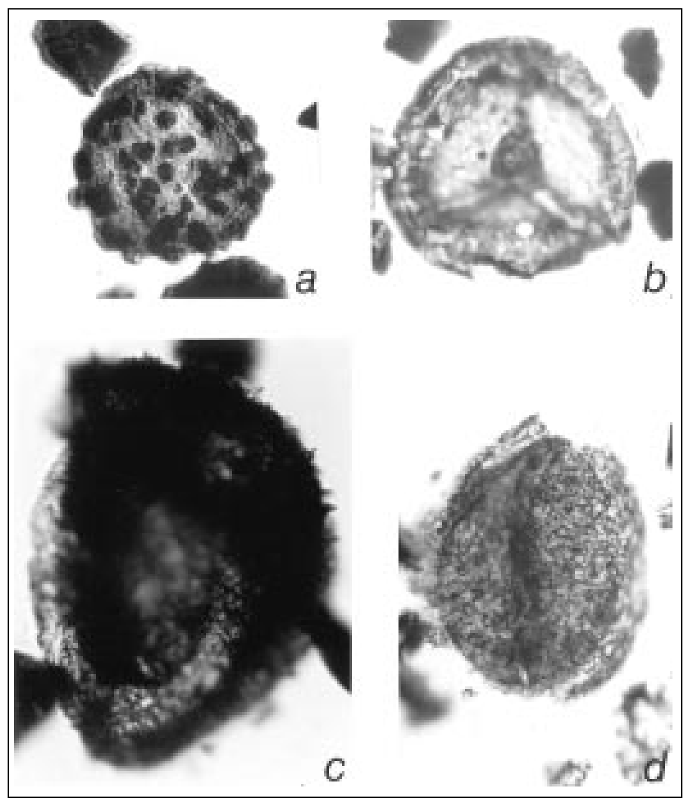

Figure 3. Photomicrographs of fossil spores from the upper Fremouw Formation, Layman Peak, and Falla Formation, Shroeder Hill area. Magnification × 760. *A. Uvaesporites verrucosus* (de Jersey) Helby in de Jersey 1971, sample AE-12/slide 5; *B. Polycingulatisporites crenulatus* Playford and Dettmann 1965, sample AE-12/slide 6; *C. Aratrisporites parvispinosus* Leschik *emend.* Playford 1965, sample LA-31/slide 1; *D. Aratrisporites wollariensis* Helby 1967, sample LA-31/slide m1.

plants in the vegetation of this part of Antarctica during the Middle and Late Triassic.

Comparison is also possible with the New Zealand Murihiku Supergroup, where marine invertebrate faunas provide good age control for the palynomorph succession (de Jersey and Raine 1990). The New Zealand Triassic, however, includes significant hiatuses, such as most of the Ladinian (upper Middle Triassic), which complicates correlations. There are also regional differences in palynomorph assemblages across the eastern Australia-New Zealand-Transantarctic Mountains sector of Gondwanaland. An example noted by de Jersey and Raine (1990) is the rarity of *A. wollariensis* in New Zealand where it is restricted to one uppermost Etalian (uppermost Anisian) sample, whereas it can be common in Australian and antarctic Lower and Middle Triassic samples. In New Zealand, *A. parvispinosus* occurs in the Kaihikuan (uppermost Ladinian) and ranges into the Otapirian (Rhaetian or top of Triassic), appearing significantly later and disappearing slightly later than in Australia. Initial examination of the Shackleton material suggests possible variation in ranges of other forms, such as some of the apiculate spores. As laboratory processing continues on the Shackleton samples, the new data may clarify regional similarities and differences in assemblages and biostratigraphic ranges across Gondwanaland.

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Permian and Triassic biogenic structures, Shackleton Glacier and Mount Weaver areas, Transantarctic Mountains

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Upper Carboniferous to Triassic stratigraphic units in the central Transantarctic Mountains were deposited in diverse nonmarine environments that record climate amelioration from glacial to temperate conditions (table). Biogenic structures are widespread and locally abundant and include discrete tracks, trails, and burrows produced by benthic animals as well as generalized sediment disruption (bioturbation). In fossil-poor units of this sequence (e.g., Pagoda, Mackellar, and Fairchild Formations), biogenic structures give crucial, if limited, information about the biota. Where plant and vertebrate fossils are abundant (Buckley and Fremouw Formations), the biogenic structures provide the only record of the bottom-dwelling fauna that played an integral role in the contemporaneous aquatic ecosystem.

During the 1995–1996 field season, biogenic structures were sampled and described from 11 locations in the Shackleton Glacier area and observed on a brief reconnaissance trip to Mount Weaver. Bioturbation on bedding planes was assessed semiquantitatively using a new field technique (Miller and Smail in press). Preliminary results highlight the contribution of the biogenic structures both to constraining the Early Permian salinity conditions and thus paleogeography of the Shackleton and Mount Weaver areas and elucidating the changes in the nonmarine fauna during this pivotal period in its evolution.

Early Permian paleogeography

The Mackellar Formation in the Beardmore Glacier area previously was interpreted as recording turbidite deposition under freshwater conditions within an inland sea that extended to the Nimrod-Byrd Glaciers area (Isbell, Seegers, and Mackenzie 1994; Miller and Collinson 1994a, pp. 215–233). Presence of marine trace fossils in the Ellsworth Mountains

Carboniferous, Permian, and Triassic stratigraphic units, rock types, and depositional environments, central Transantarctic Mountains. (From Elliot 1975, pp. 493–536; Isbell 1991, pp. 215–217; Miller and Collinson 1994a; Collinson et al. 1994.)

	Formations	Rock types	Facies/ environment
Mid-Upper Triassic Lower Triassic	(upper) (mid) Fremouw Fm. (lower)	Sandstone	Braided stream, floodplain
Upper Permian	Buckley Fm.	Sandstone, shale, coal	Braided stream, floodplain
	Fairchild Fm.	Sandstone	Braided stream
Lower Permian	Mackellar Fm.	Sandstone, shale	Lacustrine turbidite systems
Carboniferous	Pagoda Fm.	Diamictite (Tillite, shale, sandstone)	Glacial

and intervening areas suggests that the inland sea was connected to the paleo-Pacific Ocean (Collinson et al. 1994). The Mackellar Formation in the Shackleton Glacier area lacks marine trace fossils and contains known nonmarine trace fossils (e.g., *Mermia, Isopodichnus, Cochlichnus*). These data indicate that freshwater conditions persisted in the early Permian inland sea from the present-day Nimrod-Beardmore Glaciers to the Shackleton Glacier areas, implying no significant connection to the paleo-Pacific Ocean in this region (figure 1). The Mackellar equivalent in the Mount Weaver area contains densely packed specimens of small *Skolithos*, more typical of marine conditions. Its presence at Mount Weaver provides some suggestion of marine influence and connection to the paleo-Pacific Ocean (figure 2).

Evolution of nonmarine fauna

Biogenic structures in Permian formations are small (<0.5 centimeters) and restricted to fine-grained facies. We found no evidence that higher energy fluvial and turbidite channel environments were inhabited by benthic infaunal animals.

Worms and arthropods are inferred to have produced the trace fossils. Possible arthropod producers include notostracans, conchostracans, and a variety of insect nymphs. Hypothesized producers of the common trace fossils *Mermia, Helminthopsis,* and *Cochlichnus* include nematomorphs and dipteran larvae. Nonmarine aquatic faunas expanded during the late Paleozoic with aquatic insects appearing in the Permian. Holometabolous insects (including diptera) probably were not present prior to the Jurassic (Gray 1988). The abundance of several types of arthropod and worm-produced trace fossils, however, underscores the importance of benthic animals in aquatic ecosystems by the Early Permian.

Permian biogenic structures in the Shackleton Glacier area are confined to thin layers. Shallow penetration is typical of late Paleozoic aquatic biogenic structures and is consistent with the inferred domination of the benthic fauna by surface grazers and shallow burrowers.

Triassic (Fremouw Formation) biogenic structures differ from their Permian counterparts in size, depth of penetration, and facies distribution. Biogenic structures are abundant in the channel-fill sandstones of the Fremouw Formation (Collinson and Elliot 1984), although they are absent from equivalent Permian facies. Up to 30 percent of samples on sandstone bedding planes (each sample equals 0.0625 per square meter) are disrupted by bioturbation. The most common trace fossils are large (>1 centimeter), mor-

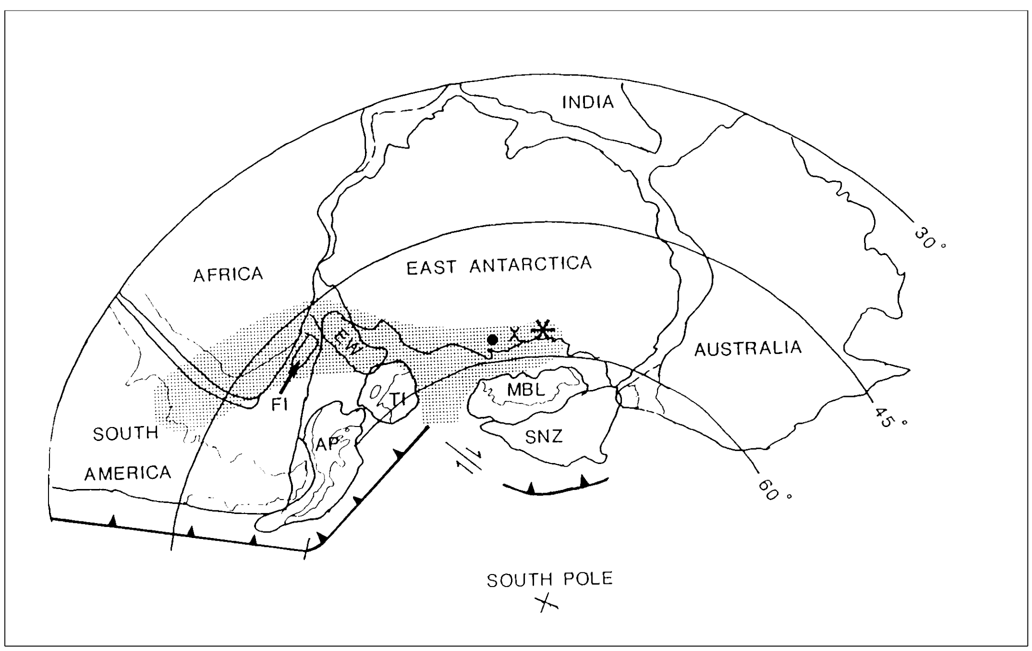

Figure 1. Reconstruction of Gondwanaland for 230 million years ago from Grunow, Kent, and Dalziel (1991) modified to show study areas relative to inland sea hypothesized to have extended over entire stippled area. (* denotes Beardmore Glacier area; x denotes Shackleton Glacier area; dot denotes Mount Weaver; MBL, Marie Byrd Land; SNZ, South Island, New Zealand; TI, Thurston Island; EW, Ellsworth Mountain block; FI, Falkland Islands; AP, Antarctic Peninsula)

Figure 2. Large oblique burrow interpreted as crayfish burrow, floodplain deposits, lower Fremouw Formation, Kitching Ridge, Shackleton Glacier area. Note light-colored, pelleted burrow walls. 15-centimeter ruler for scale.

phologically variable burrows produced by an unknown arthropod that moved deeply within the sediment (Miller and Collinson 1994b).

Large and complex burrows occurring in floodplain deposits (lower Fremouw Formation) are interpreted as produced by crayfish based on close resemblance to crayfish burrows (figure 2), which predate the previous published first occurrence of crayfish burrows (Hasiotis 1993) and provide evidence of pre-Cenozoic crayfish in the Southern Hemisphere.

Summary

Late Carboniferous to Permian biogenic structures from the Shackleton Glacier area were produced by shallow burrowing arthropods and worms in quiet-water settings within diverse nonmarine environments. Limited data do not preclude higher salinities in the Early Permian inland sea in the Mount Weaver area. By the early Triassic, channel sands were inhabited by vigorously burrowing arthropods. Crayfish burrowed deeply into the channel margins and adjacent floodplains, demonstrating behaviors similar to those of their modern counterparts.

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Preliminary analysis of Triassic vertebrates from the Shackleton Glacier region

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During the 1995–1996 austral summer, exposures of the Fremouw and Falla Formations near the Shackleton Glacier were searched for fossil vertebrates. Initially, helicopter aerial reconnaissance was used to determine the best potential sites. Localities selected for field study included Collinson Ridge, Shenk Peak, and Halfmoon Bluff in the Cumulus Hills at the junction of the Shackleton and McGregor Glaciers; Schroeder Hill and Ellis Bluff farther up the Shackleton Glacier; and Kitching Ridge and Layman Peak on the opposite side of the Shackleton.

Only the lower portion of the Fremouw Formation (Early Triassic) produced vertebrates at any of the localities. In total, 120 specimens were collected from a number of places; however, the most productive site by far was Collinson Ridge. Numerous specimens were also collected at Halfmoon Bluff, and fewer came from Kitching Ridge and Shenk Peak. Only a single specimen was found at Layman Peak. Exposures at Schroeder Hill and Ellis Bluff were both stratigraphically higher than the other sites. Field analysis indicated that only the upper part of the Fremouw Formation and the lowest part of the Falla Formation were present in these sections. Although both Schroeder Hill and Ellis Bluff had an abundance of carbonized plant material, no vertebrates were found. The upper part of the Falla Formation that had produced Early Jurassic dinosaur material near the Beardmore Glacier (Hammer and Hickerson 1994; Hammer, Hickerson, and Slaughter 1994) was not found at any site near the Shackleton Glacier. Apparently sediments that young occur only on the highest peaks in the southern portion of the Transantarctic Mountains, which are all closer to the Beardmore area.

Although preparation of the vertebrates has just begun, preliminary analysis indicates the presence of both carnivorous and herbivorous therapsids. Small skulls and skeletons of diapsids (possibly eosuchians) as well as anapsids (including at least one type of procolophonid) have also been recognized. Finally, several taxa of small temnospondyl amphibians occur, including at least one lydekkerinid and a brachyopid (see table).

Collections made in previous years from the antarctic Fremouw Formation have included the therapsids *Lystrosaurus*, *Myosaurus*, *Ericiolacerta*, *Pedaeosaurus*, *Rhigosaurus* and *Thrinaxodon*, the eosuchian *Prolacerta*, the anapsid *Procolophon*, the lydekkerinid *Cryobatrachus*, the brachyopid *Austrobrachyops*, indeterminant rhytidosteid temnospondyls, and an indeterminant thecodont (Colbert 1982, pp. 11–35; Cosgriff and Hammer 1984; Hammer 1990, pp. 42–50). Whereas some specimens collected this past season appear to belong to these previously described taxa (particularly *Lystrosaurus*), others appear to represent new genera and/or genera not previously reported from Antarctica. In particular, the abundant amount of small anapsid, diapsid, and temnospondyl material includes specimens that do not belong to known antarctic genera.

As the table illustrates, the vertebrates were collected from more than one stratigraphic horizon at Kitching Ridge, Collinson Ridge, and Halfmoon Bluff. Although previous field seasons have shown vertebrates also occur in several horizons at Shenk Peak (Cosgriff and Hammer 1982), an unusually large portion of that section was under snow cover this past

Fossil taxa collected from various Fremouw Formation localities near the Shackleton Glacier. (Numbers indicate relative position in the section; H-1 lowest horizon etc.)

Locality:	Kitching Ridge[a]		Collinson Ridge[b]			Halfmoon Bluff[c]		Shenk Peak[d]	Layman Peak[e]
Horizon:	1	2	1	2	3	1	2		
Lystrosaurus		X		X				X	
Myosaurus				X					
Thrinaxodon				X					
Cynodont genus indeterminant									
Procolophon				X			X	X	
?Procolophonid indeterminant								X	
?Eosuchian/diapsid genus indeterminant (2)				X					
Lydekkerinid genus indeterminant				X		X			X
Brachyopid genus indeterminant							X		
Temnospondlys indeterminant (2+)			X	X	X	X	X	X	

[a]H-1, green siltstone; H-2, fine-grained sandstone
[b]H-1, conglomerate; H-2, green siltstone; H-3, laminated siltstone
[c]H-1, laminated siltstone; H-2, conglomerate
[d]Conglomerate
[e]Green siltstone

season, and the snow restricted collecting to a single horizon. The table also shows that the vertebrates occur in different facies. This study will include a faunal analysis of each horizon to determine if certain depositional settings preferentially preserve certain taxa, and/or if the distribution of some taxa are related to climate.

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New observations on the Triassic stratigraphy of the Shackleton Glacier region

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During the 1995–1996 field season, we measured and described stratigraphic sections, collected paleocurrent data, and searched for vertebrate fossils at the eight Triassic localities listed in the table and located in figure 1. Several localities in the Cumulus Hills where we had worked in 1970 and 1977, including Mount Kenyon and Shenk Peak, were inaccessible because of heavy December snows.

The most interesting stratigraphic sections are in the lower Fremouw Formation on Collinson Ridge and Kitching Ridge on opposite sides of the Shackleton Glacier (*see* figure 2). A major dolerite sill follows the basal contact of the Fremouw Formation throughout the region and only thin sequences of highly baked Permian Buckley Formation are locally preserved in contact with the Triassic. Permian rocks can be identified by their carbonaceous nature and by impressions of *Glossopteris* leaves. The base of the Triassic is typically a prominent channel-form sandstone.

As noted by Hammer, Hickerson, and Collinson (*Antarctic Journal*, in this issue), vertebrate fossils occur at several

Triassic stratigraphic sections

Locality	Formation	Thickness	Age	Vertebrates
Shackleton Glacier				
Collinson Ridge	Lower Fremouw	90 m	Early	X
Halfmoon Bluff	Lower to Middle Fremouw	180 m	Early	X
Ridge on east side of Mount Rosenwald	Middle to Upper Fremouw	140 m	Early-Middle	
Schroeder Hill	Upper Fremouw	140 m	Middle	
Unnamed ridge southeast of Schroeder Hill (85°24'S 174°50'W)	Upper Fremouw-Falla	97 m	Middle-Late	
Ellis Bluff	Falla	83 m	Late	
Ramsey Glacier				
Layman Peak	Lower Fremouw	100 m+	Early	X
Unnamed ridge east of Layman Peak (84°49.5'S 179°49.5'W)	Lower to Middle to Upper Fremouw	469 m	Early-Middle	

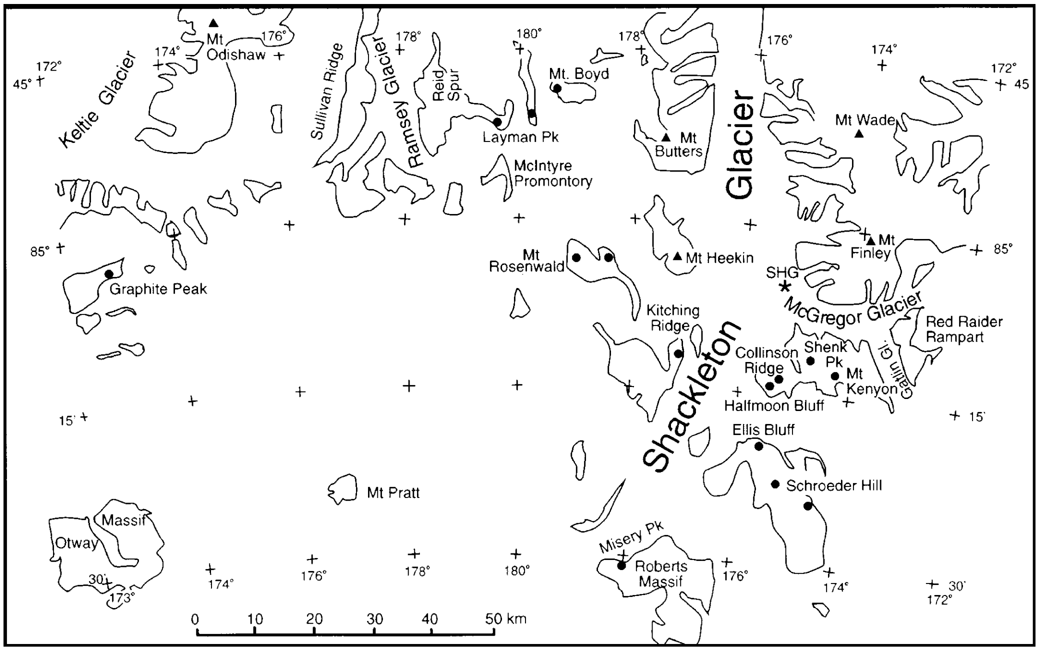

Figure 1. Map of the Shackleton Glacier region. Triassic sections measured during the 1970, 1977, and 1995 field seasons are indicated by heavy dots. SHG indicates location of 1995–1996 camp.

horizons in lower Fremouw sections (figure 2). At Kitching Ridge and Collinson Ridge, the lowest occurrences are water-worn bone fragments at the base of stream channel deposits. The bones were probably eroded out of surrounding flood-plain deposits and redeposited along with mudrock and quartz clasts downstream. Higher vertebrate horizons contain partial skeletons at the tops of major channel sandstones and within the basal part of the overlying mudrock sequences. The upper parts of the same mudrock sequences contain abundant vertical traces of roots and other paleosol indicators. The animals appear to have been buried during major floods after stream avulsion. Relatively rapid burial by flood-plain sediments, probably over weeks or days, accounts for the excellent preservation. Skeletal material is typically scattered over a small area, possibly owing to scavenging by animals before burial.

Silicified logs and upright tree stumps occur in the lower part of the sequence at Collinson Ridge. One log is 6.3 meters (m) long and gradually tapers from a diameter of 10 centimeters (cm) to 12 cm and abruptly flares to 40 cm in diameter at the base. Logs have suffered little compression from burial, suggesting early replacement by silica. Stumps are up to 0.8 m in diameter and have the thick growth rings characteristic of Permo-Triassic wood in Antarctica (Taylor and Taylor 1993). Stumps and logs that have the basal flare preserved have roots that extend laterally into the surrounding sandstone matrix, suggesting shallow root systems. In one case, a quartz clast, 4.5 cm in diameter, is imbedded in the roots of an overturned stump. Scattered quartz pebbles are common in this part of the sequence and may have been transported by uprooted trees. The sandstone, which is fine- to medium-grained and cross-bedded, represents channel bars in a braided stream sequence (Collinson, Stanley, and Vavra 1981). If the stumps are in place, the trees grew on submerged bars in the stream. Alternatively, upright stumps and logs were transported a short distance during floods.

On Collinson Ridge, a lens of silicified peat, 6 m across and 0.6 m thick, occurs within the fossil wood-bearing sequence. Cross-sections of fossil plants including a fragment of a *Dicroidium* frond were observed within this deposit. Other partly silicified coaly lenses occur in the mudrock above the wood-bearing sequence. The silicified peat, the first of its kind reported from the Lower Triassic, is being studied by Edith L. and Thomas N. Taylor at the University of Kansas.

Fossil wood and leaves are rare in rocks of Early Triassic age in Antarctica. Root traces are ubiquitous, however, in fine-grained, greenish-gray mudrock of this age, suggesting that the occurrence of fossil wood is more a problem of preservation than original tree distribution.

Vertical and horizontal trace fossils of the type described by Miller and Collinson (1994) occur in many of the sandstones (figure 3). Larger burrows near the top of the Kitching Ridge section are described by Miller and Smail (*Antarctic Journal*, in this issue). Most of these burrows are attributed to crustaceans, but at least one is large enough to have accommodated a small vertebrate.

In a previous Triassic study of this region, Collinson and Elliot (1984) conjectured a southwesterly paleoslope for the Fremouw Formation on the basis of a few readings at a few localities. In this study, the paleoslope direction was confirmed by more than 400 paleocurrent readings from several localities, most of which were toward the southwest quadrant. Readings (83) from the Upper Triassic Falla Formation were

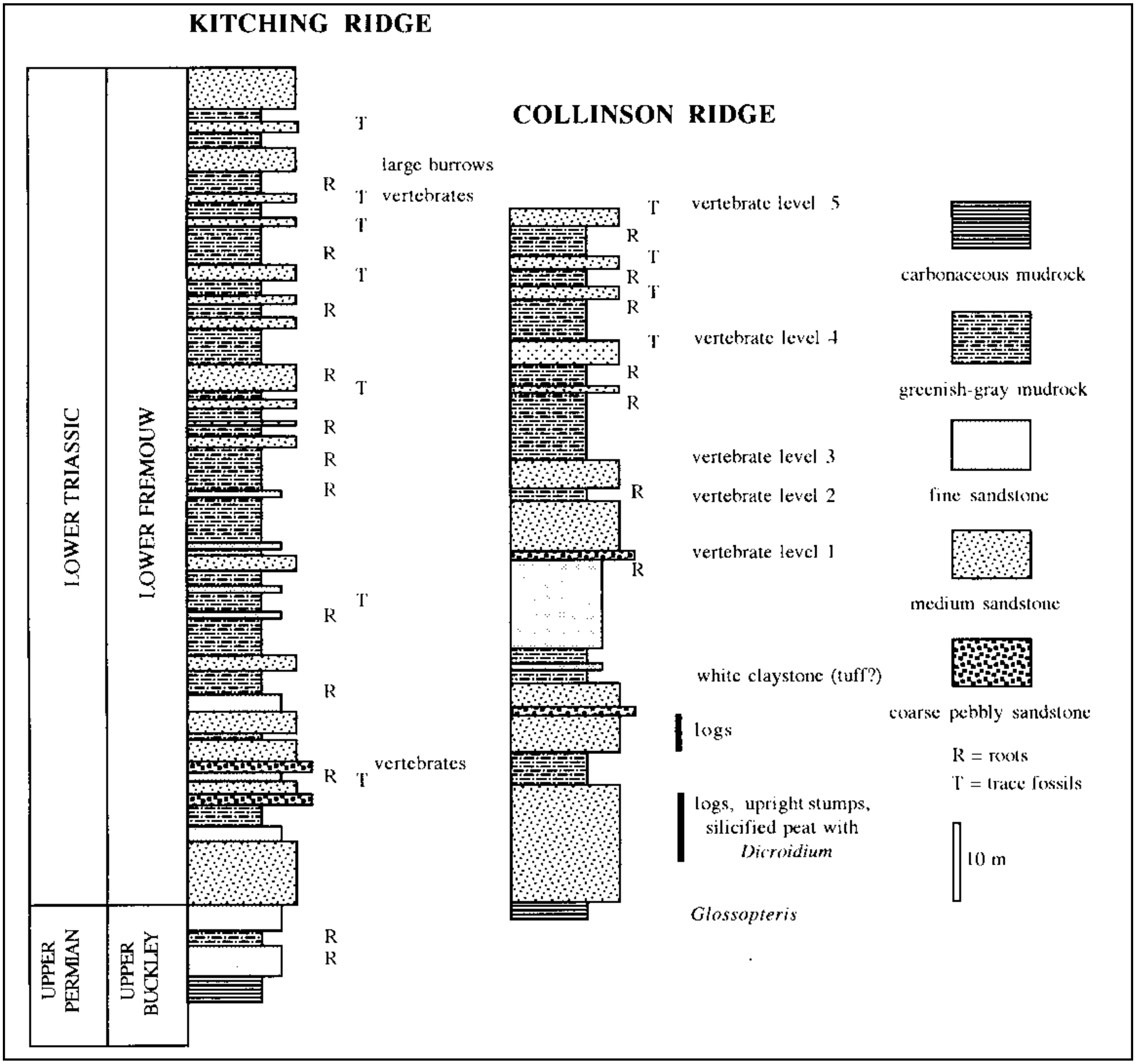

Figure 2. Lower Fremouw stratigraphic sections at Kitching Ridge and Collinson Ridge.

much less consistent and were bimodal toward the north and the south.

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Figure 3. Upright tree stump from Collinson Ridge section showing roots extending into surrounding sandstone.

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The Late Cenozoic Sirius Group of the upper Shackleton Glacier region, Transantarctic Mountains

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Examination of Late Cenozoic (Sirius Group) glacigene geology undertaken in the Beardmore Glacier region in 1985–1986 and 1990–1991 was extended to the adjacent Shackleton Glacier region during the 1995–1996 season. Our work in the latter area followed the investigations of McGregor (1965), Claridge and Campbell (1968), Mayewski (1975), and Mayewski and Goldthwait (1985) and emphasized the geographic distribution, stratigraphy, sedimentology, and structural relationships of the Sirius Group. Fieldwork was concentrated in two areas:

- the northern, or lower, Shackleton Glacier and coastal Queen Maud Mountains area, and
- the southern, or Roberts Massif-Bennett Platform area to the west and east of the upper Shackleton Glacier.

No outcrops of Sirius Group sediments were identified as a result of Twin Otter and helicopter surveys of the northern and coastal area. It had been hoped that low-elevation localities on the northern side of the Queen Maud Mountains between Shackleton and Liv Glaciers might provide glaciomarine Sirius Group successions similar to those recovered in the lower Beardmore Valley below The Cloudmaker (Webb et al. 1994, 1996a). Mayewski (1975) reported a Sirius Group locality at Mount Roth, but the outcrop was not located. Outcrops reported by previous workers at Roberts Massif and Bennett Platform were examined in detail. Minor deposits of the Sirius Group at Dismal Buttress and Half Century Nunatak were visited briefly. New localities were located at Matador Mountain, and at Schroeder Hill and Landry Bluff (Cumulus Hills). An area of approximately 1,750 square kilometers was examined in our survey of Sirius Group strata in the upper Shackleton Glacier region.

Summary of results

- The Sirius Group has been known from this area for 30 years. Our fieldwork demonstrated that its distribution is more extensive and its stratigraphy more complex than previously reported.
- The Sirius Group of the upper Shackleton Valley is interpreted mostly as the subglacial deposits of an ice stream or major

trunk glacier. Fabrics from clasts in diamictites and directions from alignment of grooves, striations, and gouge marks on the underlying dolerite trend south-north. This orientation is subparallel to the modern Shackleton Glacier trunk valley (Webb et al., *Antarctic Journal*, in this issue, 1996b).

- Stratigraphic analyses indicate that deposition entailed numerous discrete events and included both ice-contact and subaqueous modes of sedimentation.
- Glacigene successions crop out within high-relief glacial paleo-valleys on the southern or inland ice plateau side of Roberts Massif, suggesting that the Sirius Group extends much further southward beneath the present east antarctic ice sheet. This assumption gains credence from the recovery of Sirius Group sediments from inland nunataks at Mount Wisting, Mount Block, and Otway Massif (Mayewski 1975; Elliot personal communication).
- Sirius Group strata of the upper Shackleton paleovalley were originally both thicker and much more extensive, filling much of the floor of the Shackleton paleodrainage system. Large volumes of sediment were removed, probably by latest Neogene glacial processes, to expose a pre-Sirius glacial topography. The rift basins of the Ross Sea are the likely repositories of these retransported Sirius Group sediments.
- Pre-Sirius topography (the sub-Sirius Group erosion surface of Webb et al., *Antarctic Journal*, in this issue, 1996b) was relatively subdued and located at a lower elevation than today. This terrain appears to have undergone post-Jurassic–pre-Sirius fault dislocation, an episode probably associated with the early horst-graben style uplift of the Transantarctic Mountains in this region. Remobilized Sirius Group sediments were later injected as clastic dykes into fault joints in Paleozoic-Mesozoic Beacon Supergroup and Ferrar Dolerite rocks.
- Two lithostratigraphic units (formations 1 and 2 of Webb et al., *Antarctic Journal*, in this issue, 1996c) were recognized in a series of sections that crop out at Roberts Massif and Bennett Platform. In terms of the investigations of Mayewski (1975) and Mayewski and Goldthwait (1985), the sub-Sirius Group erosion surface and formations 1 and 2 are an integral part of their Queen Maud Glaciation. The lower part of our formation 1 is equivalent to their till member; the upper part of formation 1 and all of formation 2, to their stratified member.
- A major episode of post-Sirius Group structural deformation is recognized at Roberts Massif, where formerly contiguous sub-Sirius Group erosion surfaces and overlying Sirius Group sediments are separated by normal faults, which in some instances exhibit vertical dislocations of up to about 300 meters (Webb et al., *Antarctic Journal*, in this issue, 1996b).
- Mayewski and Goldthwait (1985) recognized a post-Sirius event, the Gallup Interglacial. Features associated with this event included fluvial channels and potholes cut into the surface of the Sirius Group strata at Bennett Platform and at other localities in the Transantarctic Mountains (including Mount Feather in the western Quartermain Mountains). They interpreted the Gallup Interglacial as a period of climate amelioration, surficial weathering, and abundant surface water. The surficial ferruginous weathering, rudimentary paleosols, salt layers, surface aqueous transport, erosion, and pedestal development we observed in the uppermost sediments of formation 2 at Bennett Platform are probably associated with the Gallup Interglacial.
- The youngest episode of structural deformation we observed occurs at Bennett Platform where Sirius Group successions near the edge of the platform are broken into a number of blocks or slivers and downfaulted toward Shackleton Glacier. Mayewski and Goldthwait (1985) explained deep fissures at the foot of fault scarps as fluvial channels.
- Volcanism is not associated with the pre-, syn-, or post-Sirius Group tectonic episodes recognized in the Shackleton Valley area.

Samples collected during the field season will be examined for wood, seeds, and microfossil material, notably palynomorphs, diatoms, silicoflagellates, foraminifera, insects, and vertebrate fossil debris. Pending these laboratory studies, the age of Sirius Group sediments in this area is regarded as late Cenozoic.

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The sub-Sirius Group erosion surface at Roberts Massif, upper Shackleton Glacier region, Transantarctic Mountains

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The sub-Sirius Group erosion surface is well developed and exposed on nunataks in the upper Shackleton Glacier region where a terrain of Beacon Supergroup and Ferrar Dolerite rocks have been deeply dissected prior to deposition of Sirius Group strata (Webb et al., *Antarctic Journal*, in this issue, 1996b,c). Roberts Massif provides one of the best examples of this relict landscape known from the Transantarctic Mountains and so was examined in detail during the 1995–1996 field season.

Roberts Massif

Roberts Massif, approximately 20 kilometers (km) by 20 km, is a deeply dissected and largely ice-free nunatak bordered by the inland ice plateau to the south and by the Zanneveld and Shackleton Glaciers to the east and west, respectively. Total topographic relief is a little over 700 meters (m), and Misery Peak is the highest point at about 2,723 meters. The nunatak consists of extensively faulted Beacon Supergroup and Ferrar Dolerite; the former is well exposed in the walls of deeply dissected glacial valleys, and the latter makes up extensive upland plateaus (about 2,400 m) and lowland valley floor platforms (about 2,000 m). Sirius Group sediments overlie these older rocks via a rugged disconformity, which displays buttress unconformity relationships where successions abut near-vertical paleovalley walls. Where Sirius successions occupy paleovalley floor settings, however, the disconformity surface exhibits a gently undulating relief. A pale green coloration or staining of Beacon Supergroup and Ferrar Dolerite cliff surfaces is apparent in the central region of Roberts Massif, indicating that Sirius Group strata once abutted these surfaces. These glaciated valley walls are also part of the sub-Sirius erosion surface. Deeply weathered surfaces, joints, and cracks in these older rocks have been infiltrated by remobilized Sirius Group diamictite matrices. Thick Sirius Group successions formerly blanketed much of this high-relief terrain.

A significant geomorphic feature exposed in the northern lowland region of Roberts Massif is a widespread sub-Sirius abraded pavement with an area of at least 25 square kilometers, developed on gently dipping dolerite sills. This erosion surface is one of the most extensive glacial paleosurfaces yet reported in Antarctica. The most striking aspect of the abraded pavement at Roberts Massif is the extensive development of ridges and grooves. On a scale of a few kilometers, they trend northward from near the foot of the steep scarp that stands above the northern lowland. The ridges and grooves do not appear to be controlled by either the bedrock structure or composition because the joint pattern in the dolerite is irregular and few joints are parallel to the grooves. In cross section, the pavement comprises open concave depressions and convex ridges with amplitudes of 2–3 m and wavelengths of 5–10 m. Ridges and grooves are continuous for 50–100 m, then either die out or bifurcate. The smooth abraded surface of these forms is disrupted by joints and fractures where angular blocks have become detached. Whereas vertical joint cracks measure up to several centimeters across and their interior rock walls are weathered, the later-developed sub-Sirius Group erosion surface is fresh and polished. We suggest that a phase of deep weathering preceded glacial erosion and deposition of the Sirius Group.

Sets of striations are well preserved on the ridge and groove pavement where it is protected beneath the Sirius Group strata. The main set of striations are aligned parallel to the grooves and are consistent in orientation to within 20°, i.e., north. A northerly paleoice movement direction is indicated by rare crescentic gouges and friction cracks on the ridge forms. In a few places, sets of striations with transverse orientation intersect the northerly set. These do not appear to extend beneath the Sirius Group sediments. Because a set of crescentic friction cracks was also observed on top of a thin overlying remnant of the Sirius diamictite, it is inferred that the transverse striations were produced by younger Pleistocene ice, which moved across the northern lowland during expansion of a pre-cursor of the Zanneveld and other glaciers.

What was the relationship between the Sirius Group diamictite and the underlying grooved pavement? Two-dimensional clast fabrics were measured in sediments occurring within 10 centimeters of the contact. The fabric derived is typical of a lodgement till, having a northerly preferred orientation, within 10° of that of the grooves. This finding possibly suggests that the deposition of the diamictite took place soon after the pavement was formed. Mayewski (1975) and Mayewski and Goldthwait (1985) coupled the formation of the basal erosion surface and the deposition of the overlying Sirius Group as part of a single major event, their Queen Maud Glaciation. Temporal relationships between the erosion surface and the diamictite, however, cannot be ascertained at this time.

The present lowland topography does not appear to be compatible with the ice-flow patterns derived from grooves and fabrics. One would normally expect such irregular topography to result in local variations of ice-flow directions, but this is not the case. Additionally, the pavement occurs at different levels across Roberts Massif, suggesting that the landscape was disrupted by faulting after its formation. This argument is supported by stratigraphic displacements within the Sirius Group itself. We consider that the topography during the erosion of the pavement and deposition of the Sirius Group was subdued and that the present-day high-relief terrain developed subsequently.

As defined by Sugden and John (1976, p. 194), heavily grooved surfaces of consistent orientation are generally associated with regions of areal scouring. The sub-Sirius Group erosion surface exposed at Roberts Massif is similar to the megagroove glacial topography at Kelleys Island in western Lake Erie, a feature ascribed to the passage of ice streams in a state of compressional flow (Goldthwait 1979). The orientation of grooves and fabrics described above suggests that ancestral Roberts Massif was overridden by a glacier or large ice stream flowing from the south. This ice movement does not necessarily indicate that the protoantarctic ice sheet was thicker than that of today, but rather that the Transantarctic Mountains in this area were probably less of a barrier to ice flow than is presently the case. Subsequent uplift and glacial downcutting led to the present glacial valley entrenchment of major outlet glaciers, such as the Shackleton and Beardmore.

Summary comments

- The landscape represented by the sub-Sirius Group erosion surface was developed at lower elevations and consisted of a relatively low-relief terrain of major drainage valley channels and highlands (Webb 1994). Post-Sirius Group horst-graben style normal faulting associated with Transantarctic Mountain development accentuated topographic relief and fragmented the continuity of the erosion surface datum.
- Meaningful Neogene glacial history interpretations, including ice-sheet modeling and determination of paleodrainage patterns, requires restoration of this paleolandscape to its pre-deformation configurations (Webb 1994; Webb et al. 1996a).

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Stratigraphy of the Sirius Group, upper Shackleton Glacier region, Transantarctic Mountains

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Stratigraphy of the Sirius Group

Our observations of Sirius Group stratigraphy and paleogeographic settings in the upper Shackleton Glacier region are based on outcrops in two largely ice-free areas on the western (Dismal Buttress-Bennett Platform-Matador Mountain; 250 square kilometers) and eastern (Roberts Massif; 250 square kilometers) sides of the Shackleton Glacier (Webb et al., *Antarctic Journal*, in this issue–a, –b).

The Sirius Group crops out in the northern half of Roberts Massif, at its southern margins with the inland ice plateau, and at its western margin along the Shackelton Glacier. Outcrops also occur at three localities on the western margin of the Shackleton Glacier, at Dismal Buttress, Bennett Platform, and Matador Mountain. The 3-kilometer (km) long cliffs fronting Bennett Platform provide the most complete stratigraphy of the Sirius Group observed in the upper Shackleton Glacier area. The relationship between the sub-Sirius erosion surface and basal Sirius Group sediments is best exposed at Roberts Massif (Webb et al., *Antarctic Journal*, in this issue–b), although this erosion surface is also well preserved at all localities along the western margins of the Shackleton Glacier. At Roberts Massif, this contact was observed between about 2,000 meters (m) and about 2,500 m, whereas 25 km north at Bennett Platform the contact at the base of measured sections is slightly lower at about 1,800 to about 2,000 m. Structural relationships across the 6- to 10-km-wide Shackleton Glacier have not been resolved and elevation differences for the base of Sirius Group successions might result from natural topographic contrasts, differential faulting on separate structural blocks, or some combination of the two factors.

Two distinctive glacigene stratigraphic units are recognized. We refer to these informally as formations 1 and 2. An assemblage of rocks, which may also be part of the Sirius Group, is also noted here. These are found only as erratic boulders between 1,800 and 2,000 m on the surface of Quaternary moraines around the northern flanks of Roberts Massif and along the foot of the Bennett Platform cliffs, all within a kilometer or two of the Shackleton Glacier. The boulders are up to 4 m in diameter and consist of well to moderately lithified diamictite, sandstone, and conglomerate. Bedding planes are discernible in some sandstone and conglomerate boulders. One boulder from the moraine field below Bennett Platform contained abundant wood stems and striated clasts. Clasts of these lithologies occur in formations 1 and 2 and so are older. Similar lithologies occur as large boulders on the surface of Meyer Desert in the Dominion Range. We tentatively assign these rocks to the Sirius Group. They have not been seen *in situ* anywhere in the nunataks that border the inland ice plateau and are presumed to have been derived from sub-ice-sheet regions to the south of Shackleton and Beardmore Glaciers.

Formations 1 and 2 (Bennett Platform)

Formations 1 and 2 are well exposed in cliffs below Bennett Platform. A lower unit (formation 1), which is at least 100 m thick, rests unconformably on a striated erosion surface of Ferrar Dolerite and makes up the greater part of the cliff exposure at Bennett Platform. Formation 1 includes at least eight members. Individual lithofacies present in these members include the following: massive diamictite; massive diamictites with stratified conglomerate and breccia, some of which is slumped; weakly stratified diamictites and conglomerate; stratified conglomerate and dropstone laminite; and stratified and massive diamictite, conglomerate, and laminite with common slump structures. Modes of deposition interpreted for formation 1 include the following: lodgement till, glacigenic debris flows, subaqueous bottom current reworking, glaciofluvial sedimentation, and proximal and distal glaciolacustrine deposition associated with ice-rafting and laminite sedimentation. Disruption of internal stratigraphy is also evident, especially where facies are mixed, indicating remobilization by subaqueous gravity flows. Disruption of some sequences by brittle failure is probably related to synsedimentary and/or post-Sirius Group tectonic movements.

Formation 1 and overlying formation 2 are separated by a gently sloping disconformity. Formation 2 thickens from 6 to 46 m north to south along the Bennett Platform cliff front exposures and cuts into underlying formation 1. Formation 2 contains two members, a lower massive diamictite and an overlying more friable unit of diamictite and conglomerate. This formation is not as lithified as the underlying cliff-forming formation 1 and is capped by a lag or deflation surface of dolerite boulders along the upper surface of the platform. Deep weathering, rudimentary paleosols, salt lenses, ice-wedge pseudomorphs, and rock pedestal horizons are characteristic of the upper few meters of formation 2; all phenomena were probably produced by post-Sirius Group climatic events (Webb et al., *Antarctic Journal,* in this issue–a). It is likely that the original thickness of this unit exceeded the 46 m measured at Bennett Platform. Ten kilometers to the north of Bennett Platform, near Matador Mountain, a poorly exposed succession approximately 90 m thick, is located just above glacier level and is thought to be equivalent to formation 2 at Bennett Platform. This succession suggests that relief on the intra-Sirius erosion surface between formations 1 and 2 is at least 400 m. Some part of this figure might also be ascribed to subsequent fault dislocation. The succession at Dismal Buttress is also tentatively correlated with formation 2 at Bennett Platform.

Stratigraphic correlation over the approximately 25 km separating sections at Bennett Platform and Roberts Massif cannot be made with confidence at this time. Many of the lithofacies encountered in formation 1 at Bennett Platform, including deformed and undeformed massive diamictites and laminites, are also present at several localities on Roberts Massif. At the latter locality, some diamictite successions near the base of the Sirius Group exhibit a co-occurrence of severe deformation with infra-Sirius clastic dike injection and local fault and sheared-clast zones. Formation 2 is not known at Roberts Massif.

Summary comments

- Stratigraphic analyses indicate that deposition entailed numerous discrete events and included both ice-contact and subaqueous modes of sedimentation.
- In terms of the investigations of Mayewski (1975) and Mayewski and Goldthwait (1985), the sub-Sirius Group erosion surface and our formations 1 and 2 are an integral part of their Queen Maud Glaciation. The lower part of our formation 1 is equivalent to their till member; whereas the upper part of formation 1 and all of formation 2 correlate with their stratified member.
- A careful search of the glaciofluvial and glaciolacustrine facies of formation 1, which closely resembles facies in the Meyer Desert Formation, 150 km west at Dominion Range, failed to produce plant material such as *Nothofagus* leaves and wood (Carlquist 1987; Webb and Harwood 1987, 1991, 1993, and Webb et al. 1987; McKelvey et al. 1991; Hill and Truswell 1993; Hill, Harwood, and Webb 1996; Francis and Hill in press).

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New geologic constraints on basement rocks from the Shackleton Glacier region

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During austral summer 1995–1996, Grunow, Encarnacion, and Paulsen, plus Mike Roberts, were put-in by LC-130 on 20 November to a field camp just north of Cape Surprise (figure 1, camp 1). The objective of our field programs was to collect paleomagnetic, geochronologic, paleontologic, and structural samples from basement granitoids, sedimentary, and volcanic rocks to improve understanding of the Early Paleozoic tectonic evolution of the Transantarctic Mountains. We established a Ski-doo route between Cape Surprise and the Bravo Hills for our second base camp in early December (figure 1). We encountered many large sastrugi and crevasses in the Gabbro and Bravo Hills areas making Ski-doo travel quite slow. From 9 December, our fieldwork was done by helicopter from the MacGregor camp where Bert Rowell joined us for the remainder of our season. The weather was excellent until 10 December whereafter, on most days, cloud cover obscured many of the basement exposures between Lubbock Ridge and the Ross Ice Shelf. The localities visited by Ski-doo, Twin Otter, or helicopter are shown on figure 1.

Prior knowledge of the age of basement rocks in the Shackleton Glacier area did not allow good geologic correlation with events elsewhere in Antarctica. In the field area, a thick succession of silicic volcaniclastic rocks, lava flows, and some limestones forms the Taylor Formation. It is widely correlated with the Fairweather Formation of Liv Glacier region to the east. The Henson marble forms the Fairweather Forma-

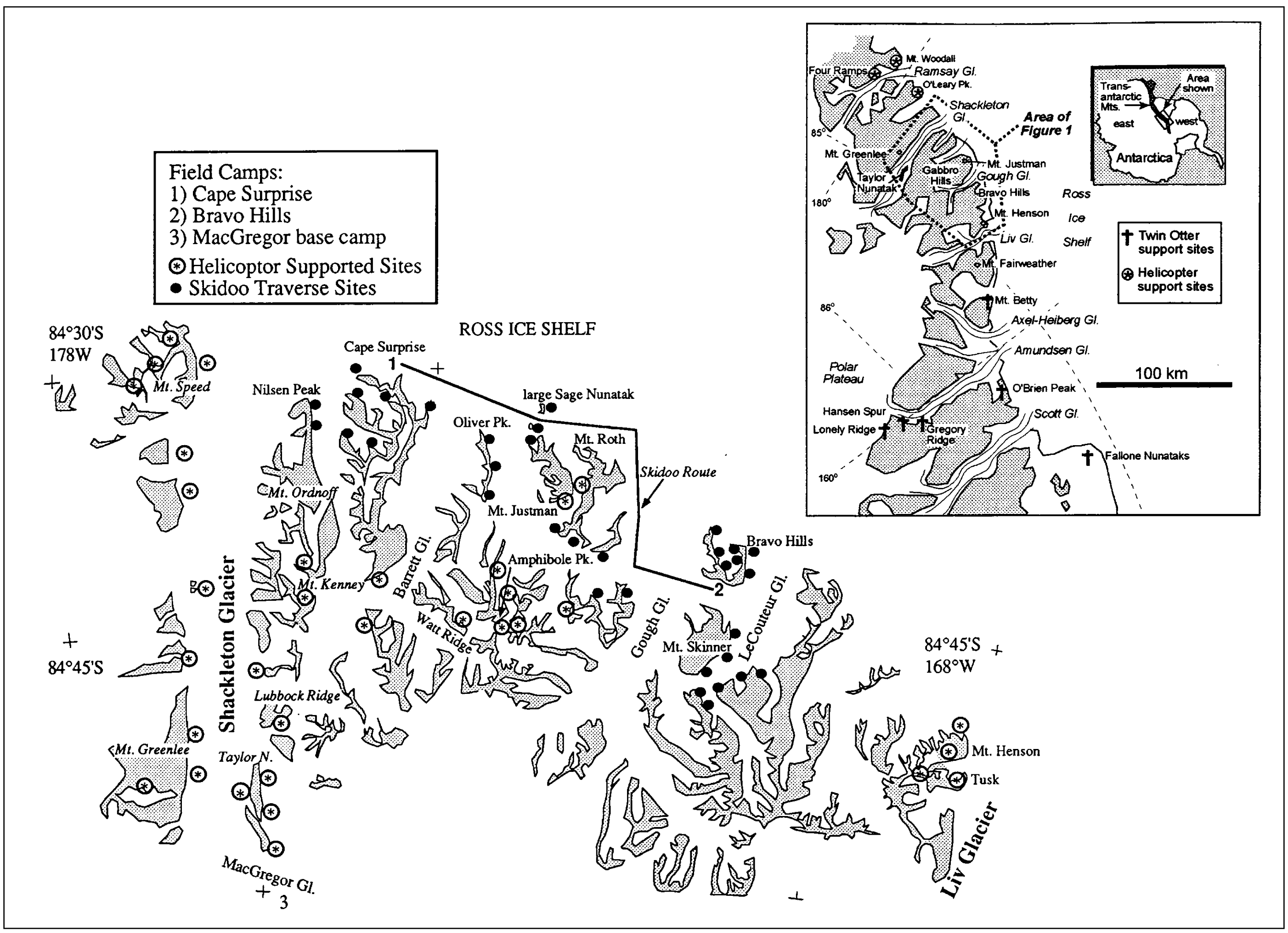

Figure 1. Sample locations and camps during the field season.

tion's upper member and was commonly regarded as the equivalent of the limestones in the Taylor Formation (Wade 1974). On lithological grounds, Wade (1974) correlated the Henson Marble with the Lower Cambrian Shackleton Limestone. One of our samples from the Henson Marble at Mount Fairweather contains what appears to be recrystallized solitary, cone-shaped, double-walled archaeocyath (figure 2).

Well-preserved trilobites from the upper part of the succession of the Taylor Formation at Taylor Nunatak reveal that the limestones containing them are late Middle Cambrian and much younger than the Henson Marble. Seemingly, the Taylor Formation is not correlative with the Fairweather Formation. The trilobites include *Amphoton* sp. cf. *A. oatesi* Palmer and Gatehouse (1972) (figure 3) and *Nelsonia* sp., cf. *N. schesis* Palmer and Gatehouse, which can be tied to Middle Cambrian successions. *Nelsonia* is endemic to Antarctica, but *N. schesis* has been reported from northern Victoria Land (Cooper and Shergold 1991, pp. 20–62), where it occurs with cosmopolitan late Middle Cambrian trilobites. This age is compatible with a 515±6-million-year uranium-lead zircon date from Taylor Formation metarhyolites on Lubbock Ridge (Van Schmus et al. in press) and suggests that the enigmatic *Cloudina?* tubes from Taylor Nunatak (*see* Stump 1995) may have limited stratigraphic value.

Structurally, the Lower Cambrian? Fairweather Formation is tightly folded, foliated, and metamorphosed to greenschist/lower amphibolite facies, whereas the late Middle Cambrian Taylor Formation is relatively unmetamorphosed and largely only tilted with no penetrative deformation. Several north-south trending subvertical shear zones, including mylonites, cut probable correlatives of the Taylor Formation along the Shackleton Glacier. These shear zones have downdip stretching lineations and may be associated with tilting of the Taylor Formation. We believe that the structural differences between the Taylor and Fairweather formations reflect structural level such that both formations were deformed during a single event. It is possible, however, that a late Early to early Middle Cambrian deformation event may have caused tight folding of the Fairweather Formation, and a second deformation event in post-late Middle Cambrian time resulted in ductile shearing and tilting of the Taylor Formation. At O'Brien Peak, a granite that intrudes deformed marbles and clastics has an S-C fabric indicating sinistral shear parallel to the mountain front. High-grade metamorphic rocks were observed at the small Sage Nunatak, Bravo Hills, Mount Woodall, and Fallone Nunataks. Most of the granitoids between the Shackleton and Liv Glaciers are undeformed except at their margins. Approximately 500 paleomagnetic drill cores and approximately 40 samples for isotopic dating were collected at the locations shown on figure 1.

Figure 2. Photograph of unprepared surface of Henson Marble member showing oblique section through a probable archaeocyath specimen, × 1. Member is from Mount Fairweather.

Figure 3. *Amphoton* sp., cf. *A. oatesi* Palmer and Gatehouse. (*Left*) A cranidium, × 3. (*Right*) An incomplete pygidium, × 8. Both specimens are from Taylor Formation, Taylor Nunatak.

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Geologic and thermochronologic studies along the front of the Transantarctic Mountains near the Shackleton and Liv Glaciers

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Since the earliest geological work in the Transantarctic Mountains (e.g., Gould 1935), it has been suggested that the Transantarctic Mountains are divided into a number of fault blocks separated by transverse structural features. Major outlet glaciers of the east antarctic ice sheet often occupy these structural features, which may represent transfer faults or accommodation zones.

The objective of this project is to study the thermal history of the Transantarctic Mountains in the Shackleton-Liv Glacier region using the fission track and argon-40/argon-39 ($^{40}Ar/^{39}Ar$) techniques. The thermal history is used to constrain the timing, amount, and rate of rock uplift and denudation. Results will be used to map the variation of these parameters across the range, between the Shackleton and Liv Glaciers, across these major outlet glaciers, and then in conjunction with other studies, map the variation along the length of the Transantarctic Mountains.

The resulting data, in conjunction with field observations, will also be used to determine the structure of the Transantarctic Mountains Front and to test the hypothesis that the mountains are segmented and that different segments have different tectonomorphological histories. It is important for constraining rift flank uplift models that not only are different denudation events in the Transantarctic Mountains identified but also that patterns of denudation across the mountains are delineated. Although the dominant uplift and denudation event that has shaped the present form of the Transantarctic Mountains began in the early Cenozoic (e.g., Gleadow and Fitzgerald 1987; Fitzgerald and Gleadow 1988; Fitzgerald 1992), the record of Cretaceous denudation events is more subtle and variable. Episodes of denudation have been recorded in the early, mid, and late Cretaceous (e.g., Stump and Fitzgerald 1992; Fitzgerald 1994, 1996, p. 35). A better understanding of Cretaceous events within the Transantarctic Mountains is critical given that most of the extension within the Ross embayment is Cretaceous in age (e.g., Lawver and Gahagan 1994) and most denudation in the Ellsworth–Whitmore Mountains crustal block is Cretaceous (Fitzgerald and Stump 1992, pp. 331–340). The little evidence we have on mid-Cretaceous denudation patterns within the Transantarctic Mountains (Fitzgerald 1994, 1995, p. 133) indicates greater denudation along the inland flank of the Transantarctic Mountains, suggesting that the subglacial basins inland of the Transantarctic Mountains are Cretaceous in age (Fitzgerald 1996, p. 35). Our sampling strategy was designed to address these questions.

Six weeks were spent in the area between the Shackleton and Liv Glaciers (figure 1), operating first out of tent camps near Cape Surprise and Mount Daniel, and then using helicopter close support from the Shackleton Glacier deep field camp (85°05.62'S 175°22.93'W). Cape Surprise is an important locality because of the unique presence of Beacon Supergroup sedimentary strata downfaulted to coastal levels (Barrett 1965). Our work at Cape Surprise (*see* Miller et al., *Antarctic Journal*, in this issue) suggests that downfaulting along the Transantarctic Mountains Front is accommodated via multiple faults rather than a single fault and that a significant component of dextral strike-slip movement as well as dip-slip movement occurred along the faults. Except for Cape Surprise, where Beacon strata and Ferrar dolerite have been downfaulted, fault location and offset, as well as the overall structure of the range, are difficult to document within the predominantly granitic basement. A series of "vertical" sampling profiles across the range (figure 2) were collected to reveal information on the denudation history, as well as the structure. Vertical profiles were collected from Mount Munson [84°48'S 174°23.3'W; 1,500 meters (m) vertical profile], Mount Olds (84°40.2'S 174°40.9'W, 900 m profile), Pyramid Peak (informal name, 84°34.3'S 174°58.6'W, 500 m profile), spot-height 700 (84°31'S 174°55'W, 500 m profile), and spot-height 950 (84°33'S 4°15'W, 550 m profile), and individual samples were collected in between these profiles at strategic locations as well as from granite near Cape Surprise (*see* figure in Miller et al., *Antarctic Journal*, in this issue). Samples were also collected in a transect from the Sage Nunataks inland along the Olliver Peak ridge system.

On the west side of the Shackleton Glacier, we collected vertical profiles at Mount Speed (84°29.8'S 176°37'W, 825 m vertical profile), Mount Wasko (84°34.1'S 176°56.3'W, 800 m vertical profile), Mount Franke (84°37'S 177°W, 1,250 m profile), and Mount Butters (84°53.5'S 177°W, 550 m profile). We anticipate this series of profiles along the Shackleton Glacier will reveal the most information related to variations in Cenozoic versus Cretaceous patterns of denudation. The Kukri Peneplain at Mount Butters, on the west side of the Shackleton Glacier, dips inland (i.e., south) at approximately 17° compared to the dip of the peneplain at Mount Munson and Mount Wade where it is approximately 2–3° to the south.

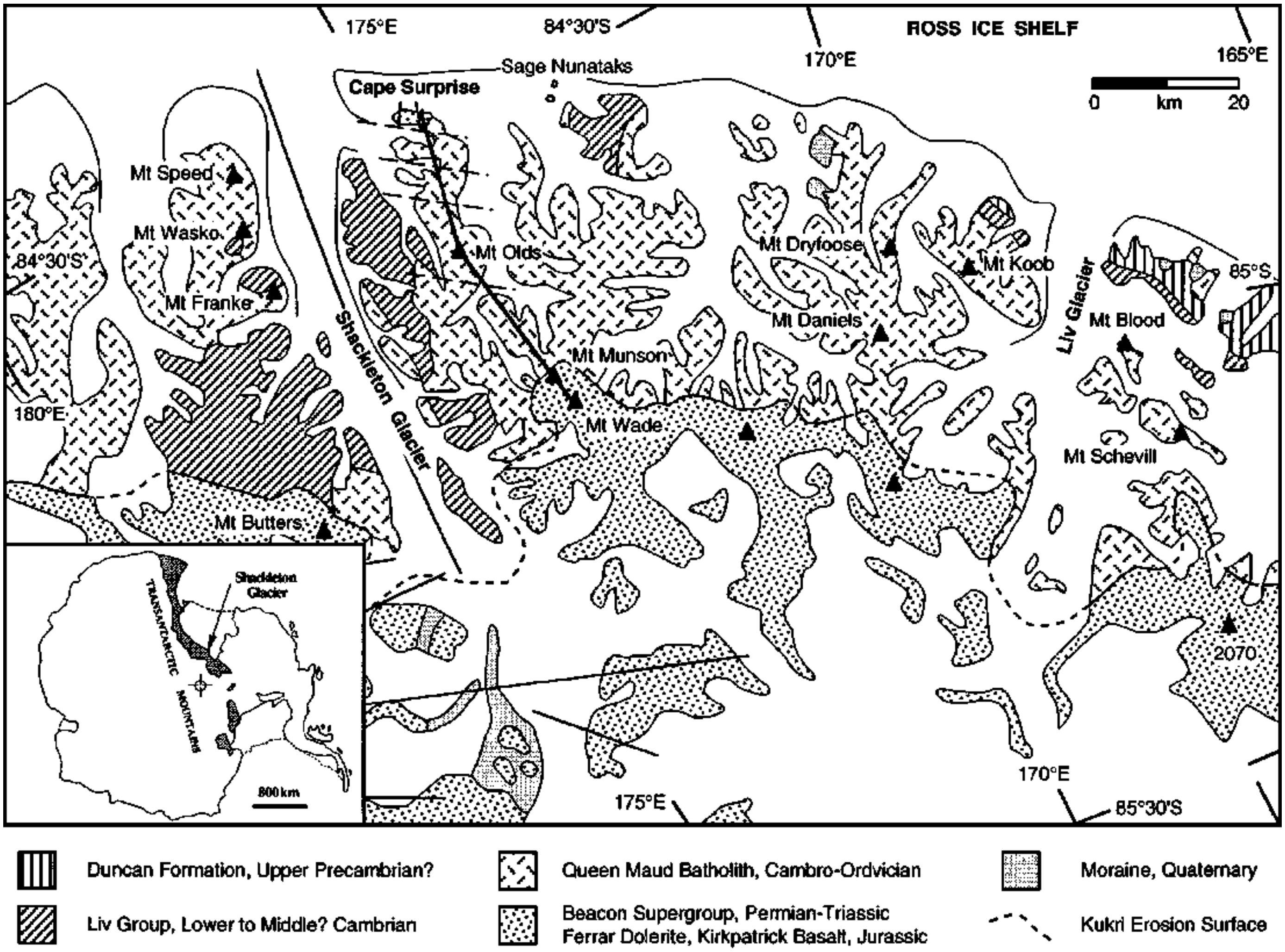

Figure 1. Geologic map of the Shackleton-Liv Glacier region (after McGregor and Wade 1969). The line between Mount Wade and Cape Surprise marks the line of the cross section in figure 2. (km denotes kilometer.)

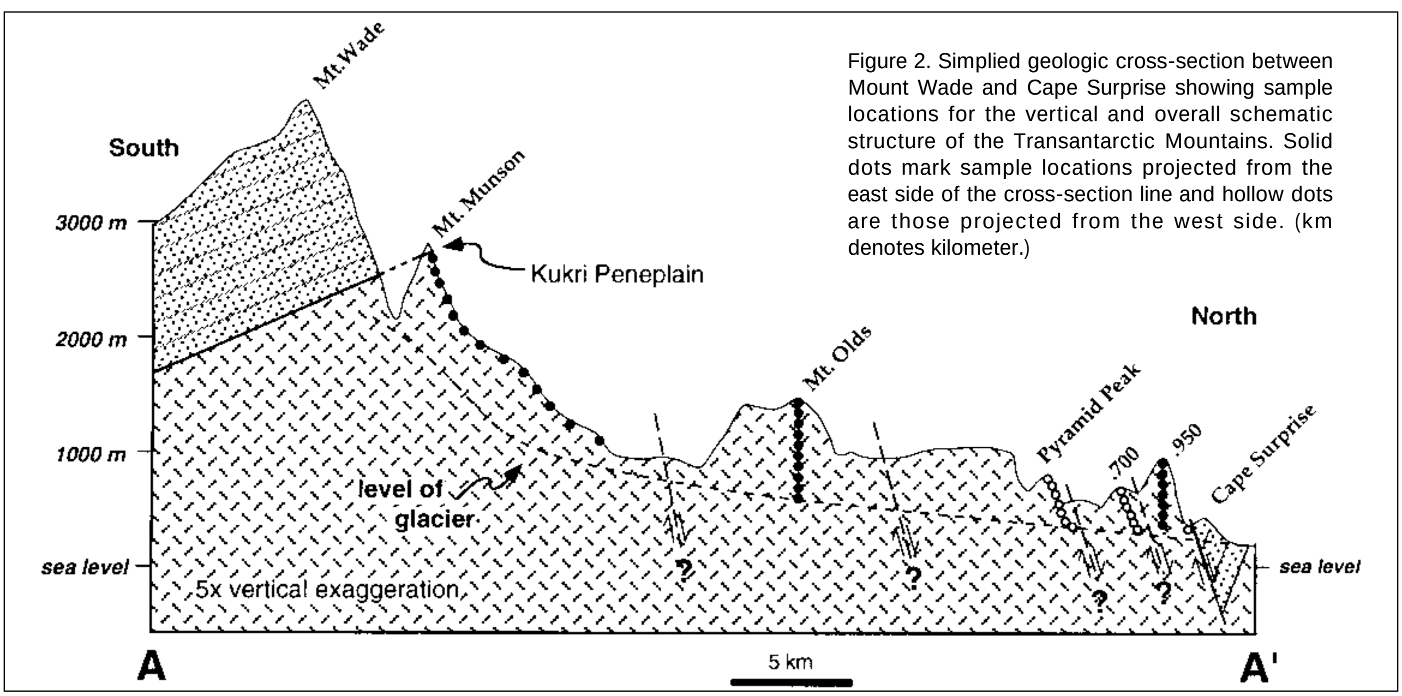

Figure 2. Simplied geologic cross-section between Mount Wade and Cape Surprise showing sample locations for the vertical and overall schematic structure of the Transantarctic Mountains. Solid dots mark sample locations projected from the east side of the cross-section line and hollow dots are those projected from the west side. (km denotes kilometer.)

These contrasting dips indicate that a transfer fault (or accommodation zone) transects the range along the Shackleton Glacier. The overall structure of the range appears to be a series of large blocks tilted south (dipping variably) both west and east of the Shackleton Glacier (figure 1).

On the west side of the Liv Glacier, we collected limited vertical profiles from Mount Daniel (84°54'S 170°W, 1,150 m vertical profile) and Mount Koob (84°48'S 174°20'W, 300 m profile) and also samples on a north-south transect along the Mount Dryfoose spur. In this area, the Kukri Peneplain can be traced from the dramatic frontal escarpment of the Prince Olav Mountains to the summit of Mount Daniel, almost halfway across the Transantarctic Mountains Front. This location for the peneplain suggests that the frontal escarpment of the mountains in this area has formed by scarp retreat rather than faulting, implying that the mountains and the frontal scarp are relatively old features, rather than having formed by relatively young tectonic uplift and faulting. Furthermore, it indicates that the Transantarctic Mountains Front occupies only the outer half of the area between the frontal escarpment and the coast.

On the east side of the Liv Glacier, we collected an approximately 800-m vertical profile by helicopter between Mount Schevill (85°5'S 167°18'W) and Mount Blood (85°1'S 167°28'W) and also a north-south transect between Duncan Mountains and the Kukri erosion surface under spot-height 2070 (85°12'S 167°6.5'W). The uniform attitude of the Kukri Peneplain on either side of the Liv Glacier does not indicate the presence of a structural feature there, although the presence of such a large glacier outlet does suggest a transverse structural feature may exist.

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Structural and geomorphological observations at Cape Surprise, Shackleton Glacier area

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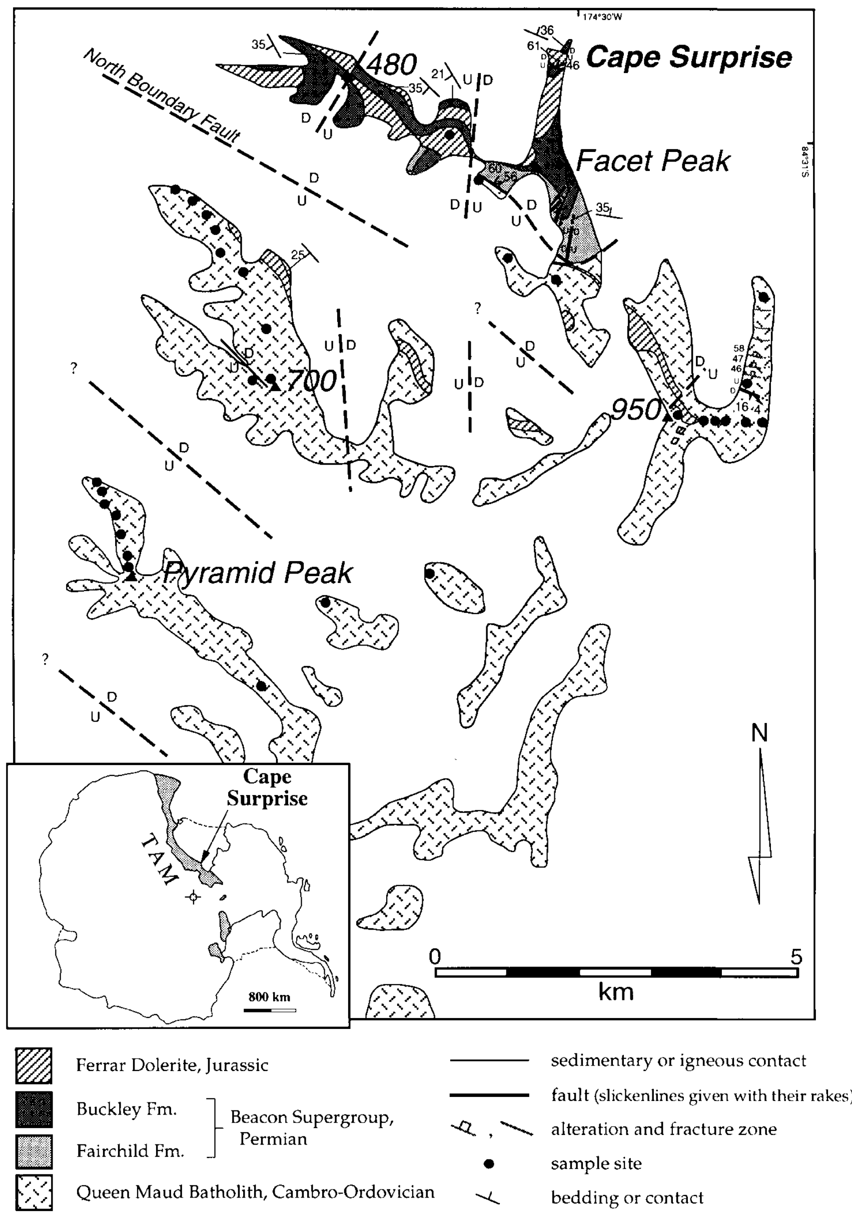
Nearly 3 weeks were spent in the Cape Surprise (84°30.27'S 176°23.3'W) region of the central Transantarctic Mountains to extend its mapped coverage farther south, to review the previously mapped geology, and to sample granitoids in a transect perpendicular to the range front for apatite fission-track thermochronology (*see* Fitzgerald et al., *Antarctic Journal*, in this issue). Cape Surprise (figure) was named by the Southern Party of the New Zealand Geological Survey Antarctic Expedition (1963–1964) because the strata of the Beacon Supergroup and sills of Ferrar Dolerite they found there were completely unexpected. Along the 3,500-kilometer (km) length of the Transantarctic Mountains, this is the only occurrence of Beacon units exposed at the coast. The closest exposure of similar Beacon strata to those at Cape Surprise is in the range's steep frontal escarpment, 2,500 meters (m) higher in elevation and approximately 35 km to the south.

In the continuing study of the geological evolution of the Transantarctic Mountains, Cape Surprise is important because it provides constraints on the structural geometry of the range and the boundary of the west antarctic rift system. Models for the formation of the present-day Transantarctic Mountains describe a relatively simple rift-flank architecture with essentially range-parallel normal faults along the coast (e.g., Fitzgerald et al. 1986; Stern and ten Brink 1989). More recent work (Fitzgerald 1992; Sutherland 1995; Wilson 1995; Fitzgerald and Stump in press), however, suggests a more complex, dextral oblique Cretaceous and Cenozoic rift margin. Because Cape Surprise is one of only two places along the Transantarctic Mountains Front (Fitzgerald 1992) where post-Jurassic normal faulting has been observed directly—the other is in southern Victoria Land—and the only place where all associated range-front offset has been proposed along a single fault (Barrett 1965), it is an important locality for models of mountain uplift that require a large range-front normal fault (e.g., McGregor 1965; Stern and ten Brink 1989).



Geologic sketch map of the Cape Surprise region (modified from Barrett 1965) showing sample locations for fission track samples. (TAM denotes Transantarctic Mountains.)

Cape Surprise comprises Permian strata of the Beacon Supergroup (Fairchild and Buckley Formations), Jurassic sills of Ferrar Dolerite, and Cambro-Ordovician granitoid of the Queen Maud Batholith (Barrett 1965; Stump 1995; Barrett personal communication). The contact between Beacon strata and granitoid on the south side of Cape Surprise was previously interpreted as a nonconformity, the Ordovician-Devonian Kukri Peneplain, by Barrett (1965) and subsequently by La Prade (1969). By extending the gently (approximately 5°) southwest-dipping Beacon beds atop Mount Wade north to the position of the cape, Barrett (1965) calculated a throw of 3,100 to 5,200 m that must have occurred to displace the units down to their present position at Cape Surprise. Without any structural data between Mount Wade and the cape, save for the presence of granitoid on Garden Spur, Barrett inferred that this offset took place across a single fault between Cape Surprise and Garden Spur, later named the "North Boundary fault" by La Prade (1969).

Our observations in the region around Cape Surprise included the following:

- The contact between Beacon strata and the granitoid southwest of Facet Peak (informal name, 84°31.2'S 174°26.8'W) is a normal fault, indicating that the Kukri Peneplain is not exposed near Cape Surprise. The well-exposed fault surface (N55°W 60°NE) (this study) is at an angle to bedding (N20°W 40°NE) (Barrett 1965). Slickenlines (raking 75°E) are present on the planar face (few square meters) of the granitoid that marks the contact, adjacent to which the sedimentary rock is brittley deformed into an approximately 20-m-wide crush zone of poorly calcite-cemented fault breccia.
- A number of other faults dissect the Cape Surprise area. Barrett (1965) inferred three: the North Boundary fault, trending west-northwest and two trending roughly north-south, cutting the western spur of the cape. In this study, west-northwest–to–northwest striking (dominantly normal dip slip) and north-to-northeast striking (including components of strike slip and rotational slip) sets of faults were observed north of Facet Peak, below spot-height 950, in the col between Facet Peak and spot-height 950, and near spot-height 700. Joints were measured in granitoid on Garden Spur, and these are divisible into similarly oriented sets.
- Fracture zones with evidence for intense hydrothermal alteration were observed, including those that cut a low spur northeast of spot-height 950 and the col between Facet Peak and spot-height 950. These zones, several centimeters to 40 m wide, often contain faults as shown by the presence of slickensides, fault breccia, gouge, offset strata, and drag folds. The trends of these zones (north-northeast and west-northwest) parallel those of other known faults in the area. Similar alteration zones have been documented by Wilson (1995) in southern Victoria Land. Craw and Findlay (1984) conclude that the hydrothermal alteration is probably the result of Jurassic Ferrar magmatism. These fracture zones have, therefore, existed with or without fault movement at least since that time.
- Ferrar Dolerite sills (approximately 80 m thick) and thinner dikes were found intruding the granitoid on Garden Spur and spot-height 950, both south of the inferred North Boundary fault. The presence of dolerite sills at these locations indicates that minimal displacement occurs between the two and that the major displacement across the North Boundary fault occurred between spot-height 950 and Facet Peak, contrary to the mapped interpretation of La Prade (1969). The basement sill must be at least 760 m below the Kukri Peneplain because granite above the sill at Garden Spur has 760 m relief. Assuming that the basement sill is no more than approximately 1,200 m below the Kukri Peneplain, as in southern Victoria Land (Hamilton et al. 1965), a revised estimate of throw across the North Boundary fault can be made: 2.4–3.1 km at its western end and 1.4–1.8 km at its eastern end. These estimates suggest that 1.3–3.8 km of throw is accommodated by normal faulting between Garden Spur and Mount Wade. To elucidate the structure of the Transantarctic Mountains on a transect between Cape Surprise and Mount Munson, we plan to apply apatite fission-track thermochronology to samples collected in systematic, vertical profiles (*see* Fitzgerald et al., *Antarctic Journal*, this issue, figure 2).
- Asymmetrical drainage patterns in the Cape Surprise area may be the result of westward fault-block tilting (*see* figure and Fitzgerald et al., *Antarctic Journal*, this issue, figure 1). Such drainage patterns are common indications of this style of deformation (Cox 1994; Keller and Pinter 1996). An angle between the strike of Beacon bedding and the inferred strike of the North Boundary fault (estimated by fault surface data and valley morphology) agrees with westward tilting, which may partially explain the east-west variance in offsets on the North Boundary fault. Block tilting at Cape Surprise and elsewhere between the Shackleton and Liv Glaciers may have been accommodated by north-south to north-east striking, east dipping normal faults underlying glaciers such as the Massam, Barrett, Gough, Le Couteur, and Morris. This west-directed drainage pattern is less distinct inland toward the Transantarctic Mountains frontal escarpment, and faults do not appear to offset Beacon strata in that face. We propose that activity on range-forming normal faults, such as the North Boundary fault, preceded block tiling along north-south to north-east striking faults and that the two episodes were separated by significant erosion and scarp retreat. Given the angles of both sets of faults to the general orientation of the range, both episodes appear related to dextral transtension in the west antarctic rift system.

We thank the National Science Foundation, Antarctic Support Associates, the Antarctic Development Squadron 6 (VXE-6), the Air National Guard, Ken Borek Air, Helicopters New Zealand, and the staff at the Shackleton Glacier camp for support during the season. This research was supported by National Science Foundation Grant OPP 93-16720. Scott R. Miller is also grateful for discussions with Robert Casavant.

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Lakes in dry valleys at 85°S near Mount Heekin, Shackleton Glacier

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Three interconnected lakes lie in a dry valley just west of Mount Heekin, which is adjacent to the middle reaches of the Shackleton Glacier (figure 1, site A). At a little more than 85°S and at an elevation of about 945 meters (m), this may be the highest and most southerly occurrence of lakes in "dry valleys" on the antarctic continent; these lakes are at a significantly higher elevation than those of southern Victoria Land. On the Mount Goodale 1:250,000 topographic map quadrangle, lakes are indicated on moraines in Moraine Canyon and at a latitude marginally to the south of those near Mount Heekin. Aerial photos suggest that these "ponds" lack stream inflow from glaciers and hence are not comparable with the Mount Heekin locality.

The uppermost lake, about 30 m across, abuts the front of an ice tongue trending northeast off the Baldwin Glacier (figure 2); when water levels are sufficiently high, this lake drains into a second lake, measuring about 150 by 65 m; the second lake drains into a third, which is about 150 m in diameter. All three lakes are ice-covered except for small moats, 1–3 m in width. The moats were open on the first visit to the valley but covered by a veneer of ice on the second visit. The ice cover on each lake is domed slightly, suggesting that the ice is frozen to the bottom. Each lake is bordered by a shoreline (visible in figure 2) rising to about 50 centimeters (cm) above lake levels at the time of observation; across these shorelines, the loose rocks at the surface are covered with dried-out algal mud, suggesting higher lake levels earlier in the season. The mud forms mats that curl as they dry out. The rocks in the moats also are covered by algal mud. Thresholds for surface flow of water from the upper to the middle lake and from the middle to the lower lake are, respectively, about 50 cm and about 90 cm high. The differences in elevation of the water levels in the moats are about 1.5 m between the upper two and about 2.5 m between the lower two. Adjacent to the shoreline of the uppermost lake is a pronounced narrow zone with gypsum efflorescence. The floor of the valley appears to consist of a coarse lag resting on a light-colored, sandy, poorly consolidated till displaying polygons as much as 10 m across and outlined by surface fractures 20–30 cm deep.

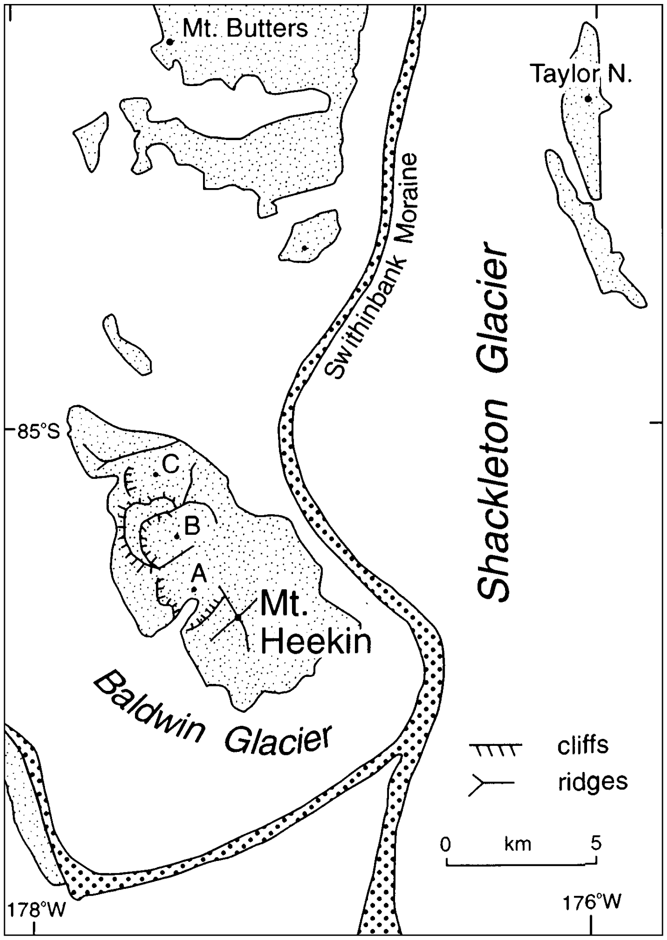

Figure 1. Sketch map to show the geographic location of the dry valleys and ice-covered lakes in the region of Mount Heekin. Areas of rock outcrop are indicated by a light dot pattern and glacial moraines by a heavy dot pattern. The three lakes occur at the site labeled A; an enclosed basin free of ice and snow is labeled B; and a third valley with a completely frozen lake is labeled C.

Unfiltered waters from the three lakes have been analyzed for major ions and nutrients (table). In general, the nitrogen chemistry is markedly different from that in the dry valleys of southern Victoria Land. Nitrate, nitrite, and ammonia are all high; the strikingly high nitrate concentrations might suggest an unusually high atmospheric input. The marked reduction, from the uppermost lake to the lowermost lake, in nitrate, nitrite, and phosphate suggests biological activity, which is, of course, supported by the occurrence of algal mats. Low calcium-to-sodium ratios suggest the possibility of relatively high input of sodium from marine sources compared to that from chemical weathering.

In the same dry valley, a slightly sinuous stream course and small gravel fan (figure 3) indicate the former presence of another lake, which lies at a lower elevation than the third ice-covered lake. The stream course connects to a small snowfield that may be the remnant of another ice tongue, which is inferred from the presence of a lobe off the Baldwin Glacier. The snowfield is too small to provide enough meltwater to maintain a lake. Drilling of this small, former lake floor might reveal an interesting history of climate changes since deglaciation.

Figure 2. Photograph showing the three lakes and the dry lake bed (indicated by an arrow) near Mount Heekin (figure 1, site A). The dry lake bed is at a lower elevation than the lowest lake. View from the northwest.

An adjacent valley to the west (figure 1, site B) is completely dry, and no signs of former lakes or stream courses were noted. Both that dry valley and the one with the lakes are closed depressions passing northeastward and northward, respectively, into a single bowl at a higher elevation; the northwestern flank of the bowl has low, wind-blown sand ridges oriented across valley.

A completely frozen lake measuring about 220 m by 65 m occurs in the next valley to the northwest (figure 1, site C), at an altitude of 1,294 m. The surface is flat, unlike the other

Selected major ion analyses for lake waters from near Mount Heekin, and for waters from the front of Wright Lower Glacier and the Onyx River, both in southern Victoria Land. All concentrations are in μM (micromolar).

Ion	Lake			Glacier front	Onyx River[a]
	Upper	Middle	Lower		
Sodium	591	223	278	17.8	169
Potassium	8.95	6.14	8.70	12.0	29.4
Magnesium	100.4	16.0	31.3	9.46	45.5
Calcium	247	31.2	70.3	32.7	119
Calcium/sodium	0.42	0.14	0.25	1.84	0.70
Ammonia	4.21	4.13	19.61	<0.15	<0.15
Nitrate	139.9	39.5	11.6	1.91	0.20
Nitrite	3.80	0.35	0.84	<0.1	<0.1
Phosphate	0.75	0.03	0.16	0.19	0.10
Dissolved inorganic carbon	257	108	336	136	313

[a]Average values for 12 samples collected during January 1996 at the weir 1 km from Lake Vanda.

lakes, but does have a series of irregular vertical cracks on the northwest flank. The ice in these cracks shows small vertically oriented columns suggestive of lake freezing. No evidence was noted of former shorelines. No glacier drains into the valley today, although a small tongue of the Baldwin Glacier probably did so in the past. Presumably the frozen lake is maintained, first, by annual snowfall and melting within the valley and, second, because the valley is narrow and, therefore, its floor is shielded from the Sun's heat by adjacent topography. Permafrost is very close to the surface upslope from the northeast end of the lake. The valley is closed off at its northeast end by a moraine with meter-size boulders.

These lakes are not obvious on air photos because of their small size, which may explain why, apparently, they were not seen and investigated on previous expeditions in the region.

Investigation of these lakes was dependent on the logistic support of U.S. Navy squadron VXE-6 and Helicopters New Zealand. Analytical data were obtained at the Crary Science and Engineering Laboratory by Dave Mikesell for anions and dissolved inorganic carbon and by Kathy Welch for cations. The x-ray determination of the gypsum salt efflorescence was made by Jeff Nicoll at Ohio State University. National Science Foundation grant OPP 94-20498 to Ohio State University provided support for David H. Elliot.

Figure 3. Photograph showing the stream course and dry lake bed indicated by the arrow in figure 2. The dry stream course is indicated by a series of arrows and the dry lake bed by an arrow labeled LB. View from the northeast.

Seed fern reproductive organs from the Shackleton Glacier area

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One of the most fascinating groups of fossil plants are the so-called Mesozoic seed ferns, a loose collection of principally impression/compression remains characterized by seeds produced in cupules. Although some of the vegetative and reproductive organs of each of the four orders (Caytoniales, Peltaspermales, Corystospermales, and Petriellales) are reasonably well documented, there is little agreement as to the origin(s) of the group and even less consensus as to their evolutionary relationships with modern seed plants. Since the initial description of the group by Thomas (1925), numerous authors have offered various interpretations regarding the nature of the reproductive organs, especially the cupule. This organ has been the focal point in most discussions that relate the Mesozoic seed fern groups with the flowering plants, because it has been variously interpreted as the precursor or homologue of the angiosperm carpel (e.g., Rothwell and Serbert 1994 and references therein). Even within the Petriellales, a group based on permineralized cupules from the Triassic of Antarctica (Taylor, Del Fueyo, and Taylor 1994), little evidence suggests a relation-

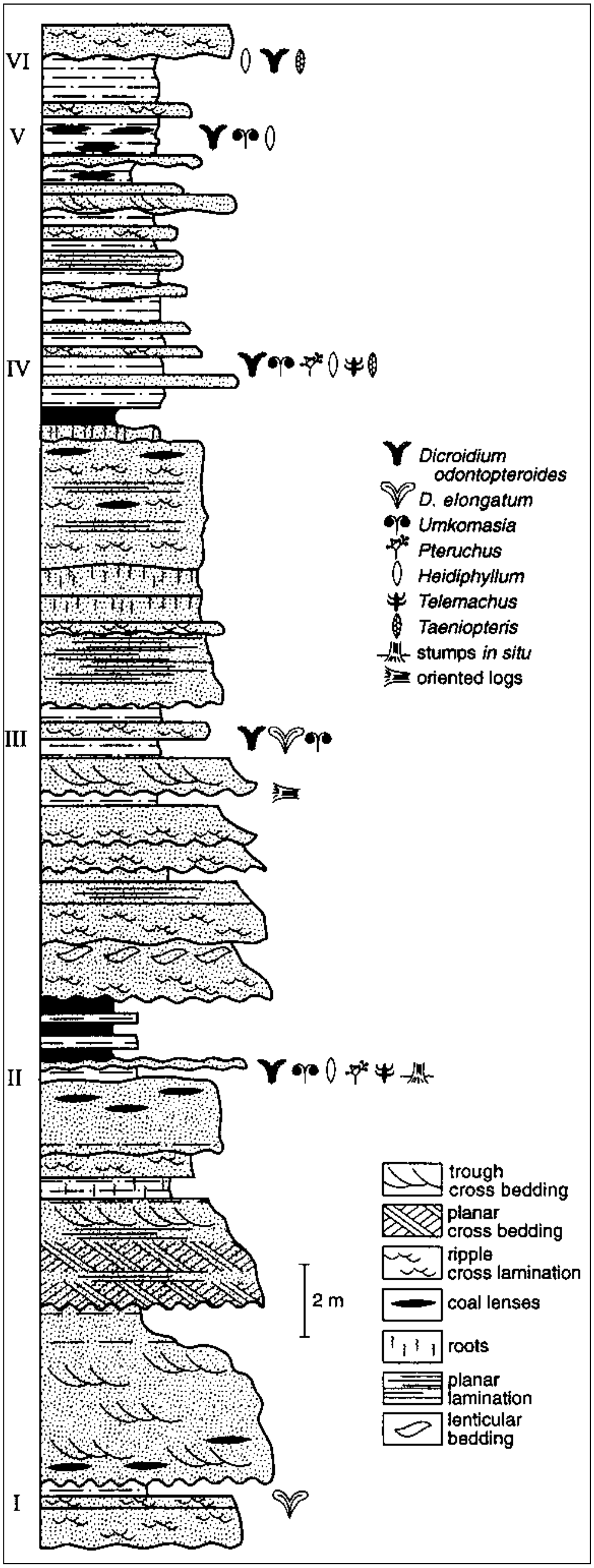

Figure 1. Section at "Alfie's elbow" locality showing plant levels.

ship between the cupule and carpel because the cupules of *Petriellaea* demonstrate a folding pattern that is different from the conduplicately folded carpel of flowering plants. In addition to differing opinions as to the nature of the seed-bearing reproductive organs, even less information is available about how and where these structures were borne and the overall habitat of the plant.

Fieldwork during the 1995–1996 season in the Shackleton Glacier region uncovered an exceptional outcrop of the Upper Fremouw–Falla Formation containing well-preserved impression/compression plant remains and a few fragments of silicified peat. The section is approximately 42 meters thick and includes six plant levels of late middle to early late Triassic age (figure 1). The site occurs on "Alfie's elbow," an unnamed ridge southeast of Schroeder Hill at approximately 85°23'717"S 174°49'916"W. The rocks at this locality represent fluvial deposits that include several cycles of channel, levee, floodplain, crevasse splay, and swamp facies. A few paleosols present are associated with the floodplain facies. It appears that the fluvial system at this site was dominated by low-energy braided streams, allowing for the preservation of fine-grained siltstones and shales that contain well-preserved plant remains, including delicate reproductive organs attached to the axes that bore them.

Of particular significance is the specimen illustrated in figures 2 and 3. It consists of a stem approximately 1.5 centimeters in diameter from which arise evenly spaced, short lateral axes up to 3 centimeters long. On the surface of these

Figure 2. Stem bearing truncated axes (arrows) each of which terminates in several reproductive organs. ×0.85.

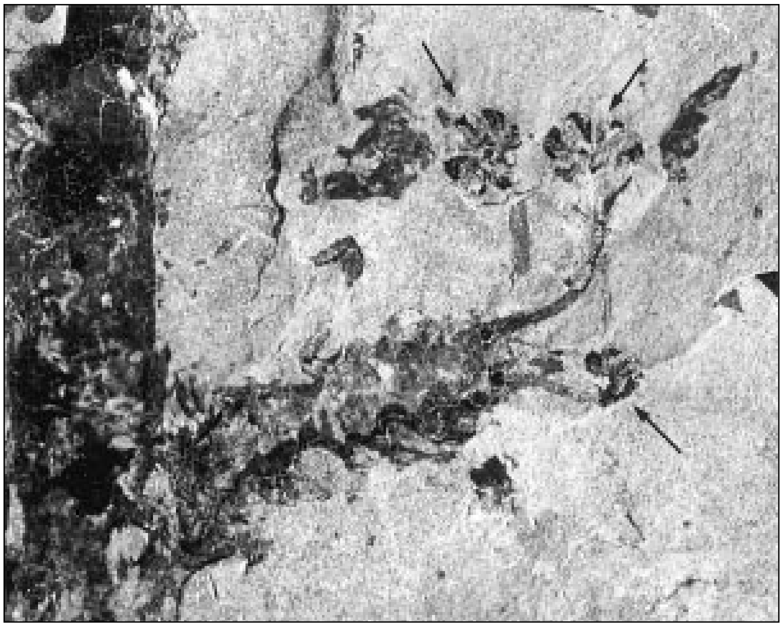

Figure 3. Detail of reproductive organs (arrows). ×1.5.

axes are closely spaced lenticular scars that may represent the former position of leaves. Attached to the tip of some of these branches are elongate slender axes each of which terminates in a cupule. Each cupule is flattened and appears to consist of several units that correspond to either cupule lobes or seeds. Present at the same site as the attached cupules are leaves of *Heidiphyllum*, a form genus of elongate, broadly sessile leaves with parallel veins that show some anastomoses. Anderson and Anderson (1989) suggest that *Heidiphyllum* is the leaf of the plant that produced the coniferlike seed cone *Telemachus*. We speculate that *Heidiphyllum* leaves may have been produced by the seed fern that bore the cupulate organs described above. We are unaware, however, of any Mesozoic seed fern cupulate structures that are attached to delicate petiolelike axes like those from the Shackleton site. Morphologically, these reproductive axes appear similar to the short or spur shoots of some ginkgophytes, a group that was well represented by Triassic time. We are hopeful that the Shackleton specimens will reveal sufficient detail so that the affinities of this interesting fossil can be determined.

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Permian and Triassic paleosols and paleoenvironments of the central Transantarctic Mountains, Antarctica

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We embarked upon our study of paleosols as guides to Permian and Triassic paleoenvironments around the Beardmore and Shackleton Glaciers of the central Transantarctic Mountains with some trepidation because few fossil soils had been reported there (Horner and Krissek 1991). It was pleasing to discover abundant paleosols (341 of 16 different kinds) in a stratigraphic section of 568 meters (m) spanning the Permian-Triassic boundary on a low northwestern spur of Graphite Peak (figures 1 and 2; 85°2.99'S 172°21.65'E) and also 34 of 6 kinds within 40 m measured in the natural amphitheater low on the northeast face of Mount Rosenwald (figure 3; 85°3.09'S 178°29.52'E). Each pedotype was named after antarctic geologists: Alton Wade, Kerby LaPrade, James Morton Schopf, James Waller Collinson, David H. Elliot, William Roy Hammer, Julia Miller, and Molly Miller. Other names were carried through from paleosols in southern Victoria Land (Retallack, Krull, and Robinson 1996). Each pedotype represents a unique ancient ecosystem.

Coals are the most obvious paleosol horizons of the Late Permian, Buckley Formation. Two kinds of paleosols bear coal: the James pedotype has impure coal and subhori-

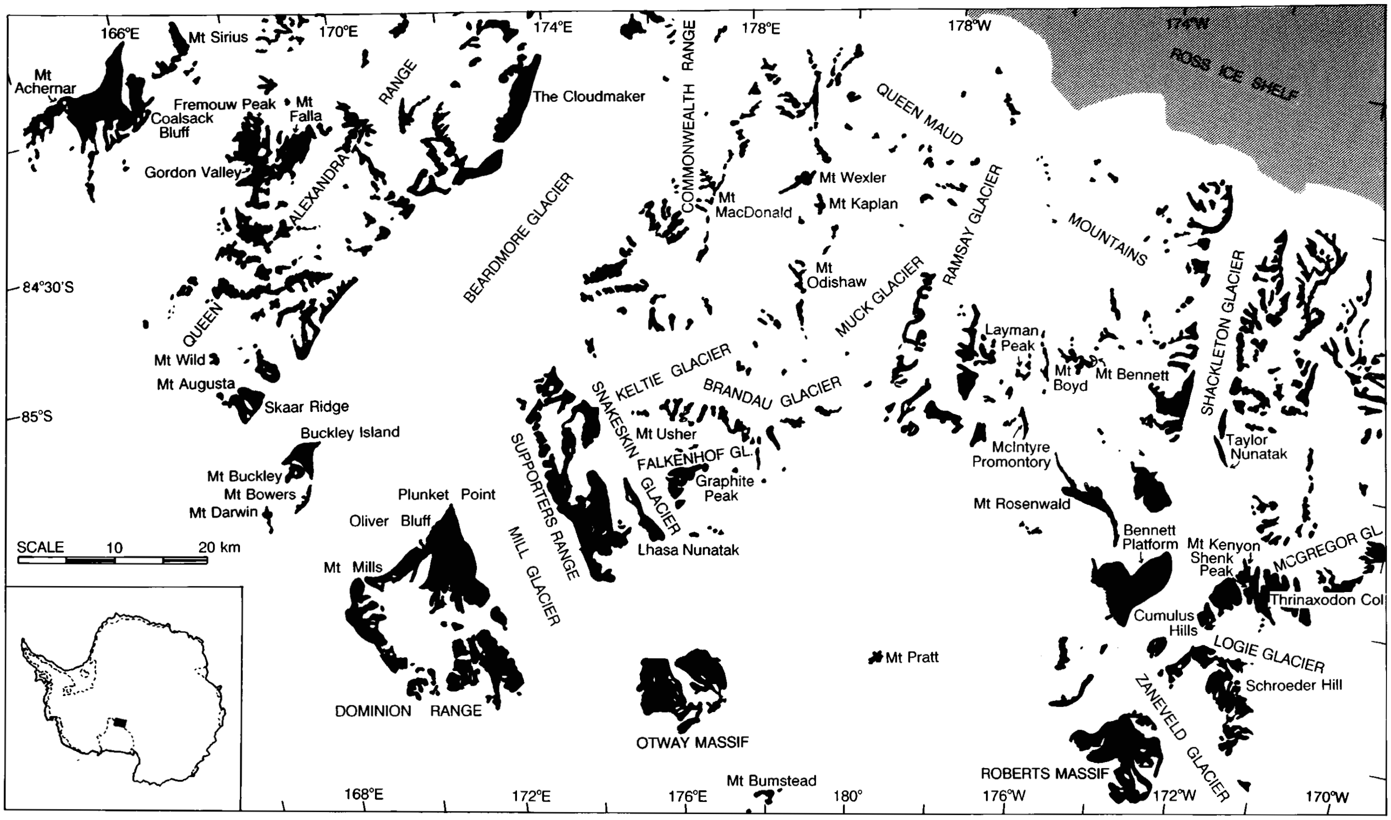

Figure 1. Rock and soil (black) and locations of Graphite Peak and Mount Rosenwald in the central Transantarctic Mountains, Antarctica. (km denotes kilometer.)

zontal *Vertebraria* roots in the underclay, but the Evelyn pedotype has bright coal and *Vertebraria* penetrating deeply [more than 5 centimeters (cm) and up to 92 cm]. Chambered roots as well as fossil stumps and leaves of *Glossopteris* are evidence that James and Evelyn pedotypes supported swamp woodlands. Their differences reflect reduced sediment supply and an initially deeper water table in Evelyn compared with James paleosols. Cherty tuffaceous siltstones and claystones include two kinds of very weakly developed paleosols. Morton pedotypes have blue hue, *Glossopteris*, thin *Vertebraria*, and rare burrows, but Molly pedotypes have green hue, *Paracalamites*, thin unchambered roots, and locally common burrows. Both represent communities early in ecological succession from disturbance. Other colonizing communities are represented by sandstones with relict bedding and root traces, such as the Sandy (gray with carbonaceous root traces) and Kerby pedotypes (greenish gray with clayey root traces). Finally, the Waller pedotype is siltstone with abundant burrows and nonchambered, gymnospermous root traces. Sandy, Kerby, and Waller pedotypes supported woody gymnospermous vegetation but lacked identifiable fossil plants. The abundance of cherty and coaly pedotypes in the Buckley Formation at Graphite Peak is remarkably similar to paleosols of the Late Permian, Newcastle Coal Measures in southeastern Australia.

Paleosols of the Early to Middle Triassic Fremouw Formation are of entirely different pedotypes. They also are more even in their development and distribution than Permian paleosols (figure 2), indicating a marked paleoenvironmental change across the Permian-Triassic boundary. Some Early Triassic paleosols of the lower Fremouw Formation are identical to those in the Feather Conglomerate and Lashly Formation of southern Victoria Land. These include John (gray and thick with subsurface accumulation of clay), Dolores (gray and thin with subsurface nodules of berthierine), Michael (lithic sandstone with stout white root traces), and Edwin (quartz sandstone with burrows and root traces) pedotypes. These can be interpreted, respectively, as soils of lowland forest, swale woodland, streamside colonizing woodland, and channel bar herbaceous vegetation (Retallack et al. 1996). Also found were William (carbonaceous shaley weakly bedded surface) and Alton pedotypes (densely rooted carbonaceous surface horizon over light-colored subsurface). Both supported lowland seasonally waterlogged woodlands that were colonizing and well established, respectively. These noncalcareous paleosols indicate humid climate persisting into the early Triassic, like similar paleosols of the lower Narrabeen Group of the Sydney Basin, Australia.

Red paleosols are a distinctive feature of the Fremouw Formation in the Shackleton and Beardmore Glacier area. At Graphite Peak (figures 1 and 2), they are 50–112 m above the base of the formation. Red paleosols were also found at a comparable stratigraphic level at Mount Rosenwald (figure 3), on the 2,870-m peak near Mount Layman (84°48.70'S

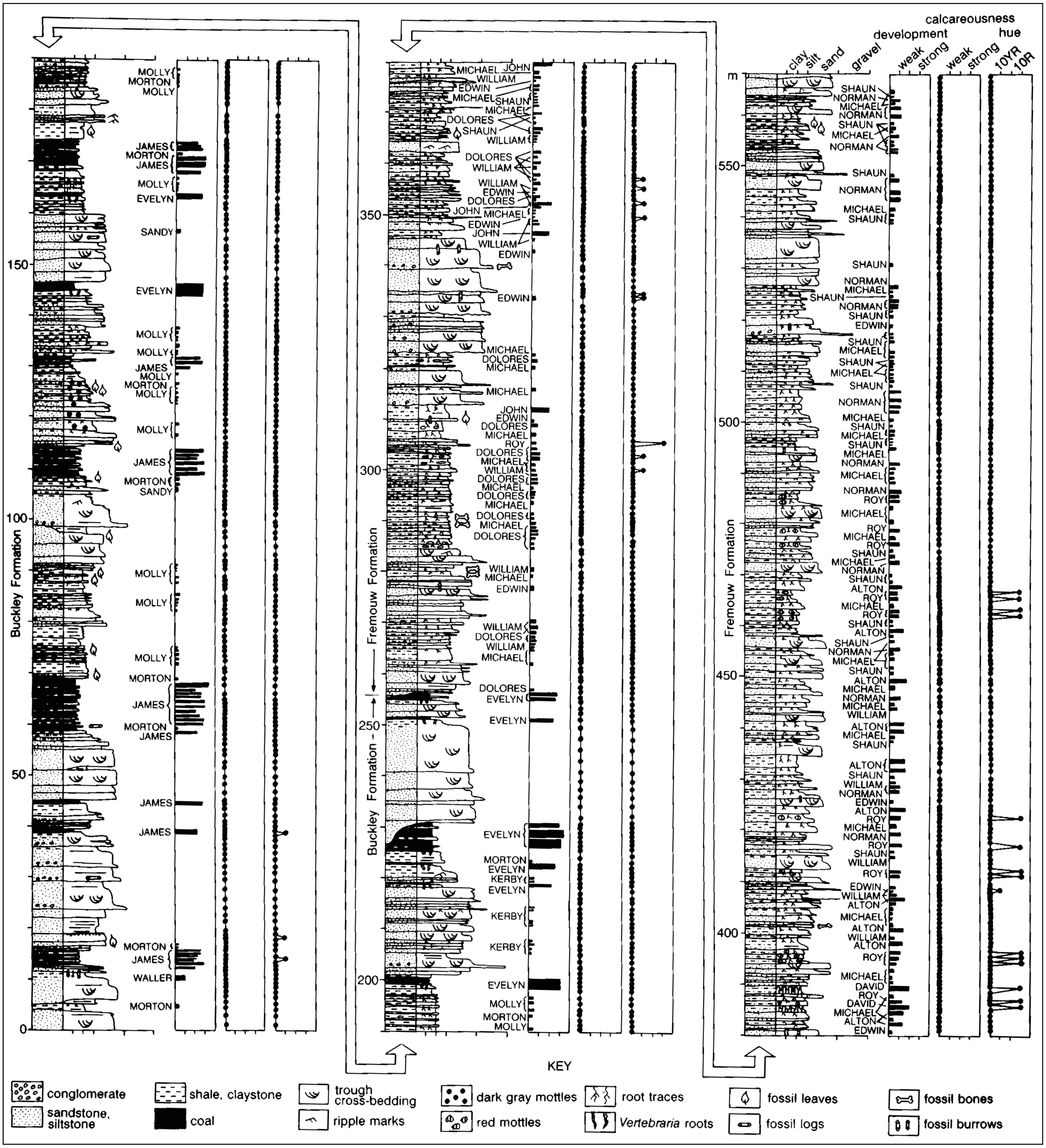

Figure 2. A measured section of paleosols on Graphite Peak. Hue is from a Munsell color chart. Calcareousness is by reaction with acid. Degree of development is from relative destruction of bedding.

179°78.18'W) and on Mount Boyd (84°48.83'S 179°68'W). The red paleosols included Roy (shaley with scattered red mottles), Julia (sandy with red subsurface), and David pedotypes (with red clayey subsurface). All have abundant stout drab-haloed root traces and are comparable to humid, cool-temperate, forested paleosols of the middle to upper Narrabeen Group of the Sydney Basin, Australia (Retallack 1977a).

Also as in the Sydney Basin, paleosols higher in the Fremouw Formation associated with Middle Triassic floras (Retallack 1977b) are gray and green. These Middle Triassic paleosols include Shaun (shale with root traces), Michael

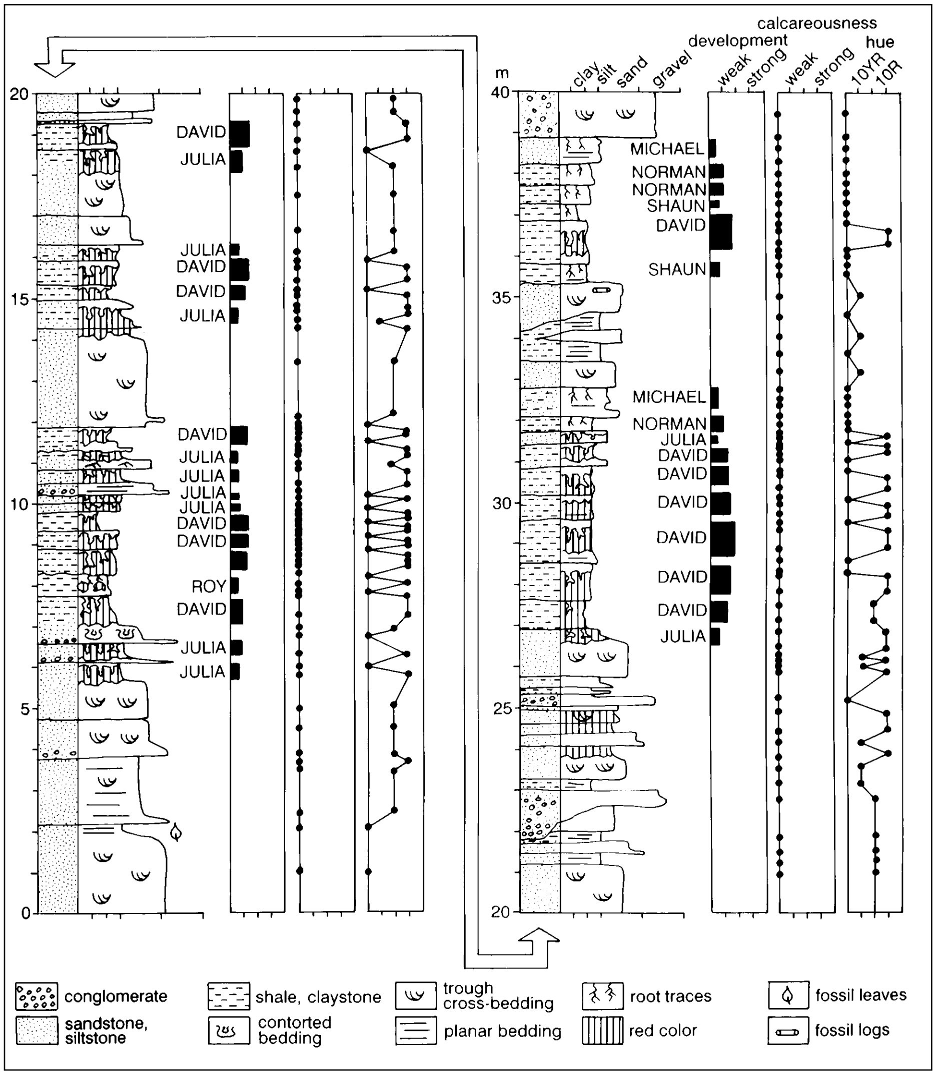

Figure 3. A measured section of paleosols on Mount Rosenwald. Hue is from a Munsell color chart. Calcareousness is by reaction with acid. Degree of development is from relative destruction of bedding.

(sandstone with root traces), and Norman pedotypes (green claystone with white root traces), identical to paleosols of the Lashly Formation in southern Victoria Land. They represent seasonally waterlogged, subhumid, cold-temperate woodlands and early successional streamside vegetation (Retallack et al. 1996).

We thank Shaun Norman for field assistance, and in Shackleton base camp of 1995–1996, we appreciated the organization of David Elliot, Kevin Killilea, and the pilots of Helicopters New Zealand. Work was funded by National Science Foundation grant OPP 93-15228.

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Paleoenvironment of the Triassic therapsid *Lystrosaurus* in the central Transantarctic Mountains, Antarctica

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Graphite Peak was the first place in Antarctica where Triassic fossil vertebrates were discovered. In 1967, Peter Barrett found a fragment (Barrett, Baillie, and Colbert 1968) that was later identified as the labyrinthodont *Austrobrachyops jenseni* (Colbert and Cosgriff 1974). Since then, Graphite Peak has yielded other fossil vertebrates of the Early Triassic *Lystrosaurus* fauna, including the herbivorous tusked therapsid *Lystrosaurus murrayi*, the small carnivorous therapsid *Thrinaxodon liorhinus*, and the eosuchian *Prolacerta broomi* (Colbert 1974; Colbert and Kitching 1977; Hammer 1989, 1990). *Lystrosaurus* was the dominant taxon of a cosmopolitan fauna of low diversity that survived the great extinctions at the end of the Permian (Cosgriff, Hammer, and Ryan 1982; Retallack 1995).

The habitats of *Lystrosaurus* in Antarctica have remained enigmatic because the majority of these fossils were collected from gravelly portions of ancient stream sandstones, where the bones had been transported some distance from where the animals lived (Colbert 1974). Also transported were skeletons of *Thrinaxodon liorhinus* in the flaggy upper portion of a paleochannel on Thrinaxodon col near the confluence of the Shackleton and McGregor Glaciers (locality described by Collinson and Elliot 1984; and visited by us). In contrast, discovery of a skeleton of *Lystrosaurus* on the buried soil on which it died allows a clearer idea of the paleoenvironment of this ancient reptile. These remains are in the saddle between two prominent sandstone bluffs on a low northwestern spur of Graphite Peak (85°2.99'S 172°21.65'E). Much of the skeleton is spread out over about a meter of the bedding plane (figures 1 and 2).

The skeleton is not articulated as an animal overwhelmed by catastrophic burial in a stream would be (figures 1 and 2). Instead, the bones are scattered and weathered, as if the corpse rotted on a soil and then fell apart. The bones were preserved partly in a hard green nodule that forms the lower horizon of an overlying soil similar to the one on which the skeleton lies. This kind of paleosol is common in the lower Fremouw Formation at Graphite Peak (figure 3) and in the Feather Conglomerate of southern Victoria Land, where it has been called the Dolores pedotype (Retallack, Krull, and Robinson 1995). Its drab color with berthierine nodules is evidence of a lowland seasonally waterlogged environment, and its weak development indicates soil formation for only a few hundred to a few thousand years. Fossil horsetails have been

Figure 1. Skeleton of *Lystrosaurus murrayi* in the lower Fremouw Formation at Graphite Peak, central Transantarctic Mountains, Antarctica.

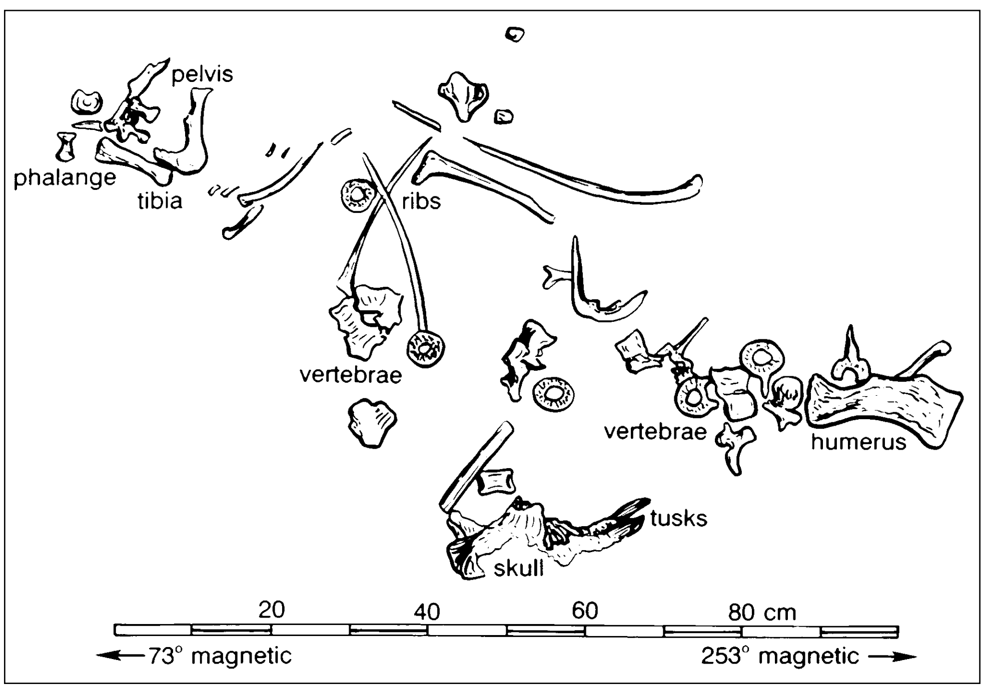

Figure 2. Scale drawing of skeleton of *Lystrosaurus murrayi* in the lower Fremouw Formation at Graphite Peak, central Transantarctic Mountains, Antarctica. (cm denotes centimeter.)

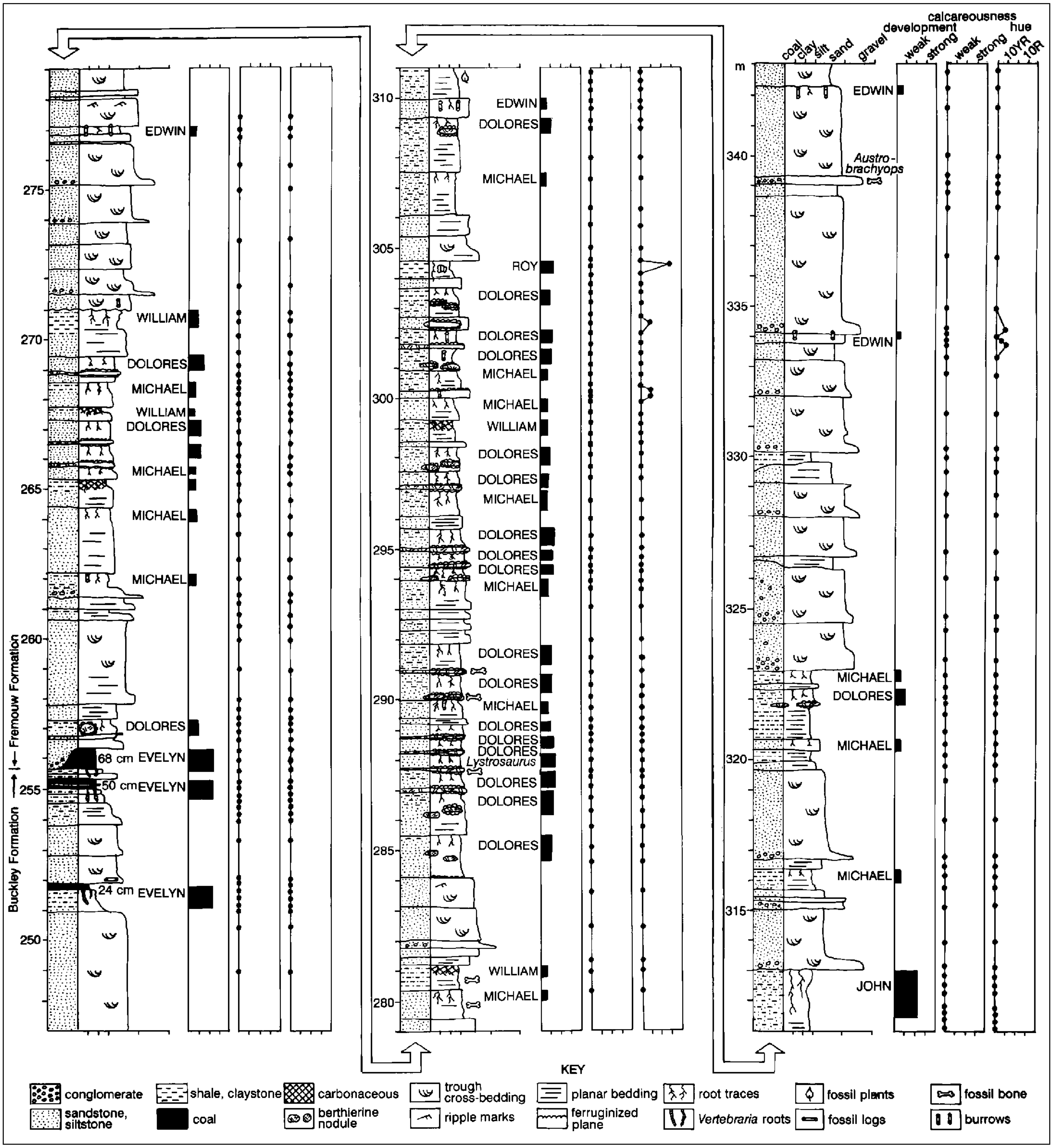

Figure 3. Location of the skeleton of *Lystrosaurus* and other vertebrate fossils in a measured section of paleosols on Graphite Peak. Hue is from a Munsell color chart. Calcareousness is from reaction with acid. Development is from relative destruction of bedding.

found in similar paleosols in southern Victoria Land (Retallack et al. 1995), but these would have been an understory to trees represented by common woody root traces. Dolores paleosols probably supported lowland colonizing woodland among forests represented by other pedotypes (figure 3).

These observations dispel a persistent notion that *Lystrosaurus* was strictly aquatic or amphibious. They support instead indications from functional morphology (King and Cluver 1991), that *Lystrosaurus* was fully terrestrial and capable of burrowing. A nonaquatic lifestyle for *Lystrosaurus* is also supported by the absence of webbing between the toes on all plausible footprints (Watson 1960; MacDonald, Isbell, and Hammer 1991; Retallack in press). Paleosols can now be added as evidence for the habitats of *Lystrosaurus*.

Shaun Norman, Evelyn Krull, and Scott Robinson assisted in the field. The organizational efforts of David Elliot and Kevin Killilea and the skilled pilots of Helicopters New Zealand in the Shackleton base camp of 1995–1996 were also appreciated. Work was funded by National Science Foundation grant OPP 93-15228.

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Late Triassic hummocky coals near Schroeder Hill, central Transantarctic Mountains, Antarctica

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Hummocky and cracked coal seams can be useful paleoclimatic indicators for cool climates (Krull and Retallack 1995). This paper discusses a coal seam with marked thickness variation, 10 kilometers southeast of Schroeder Hill, central Transantarctic Mountains (85°23.741'S 174°50.208'W). These strata are in the Falla Formation (Collinson and Elliot 1984), which has been dated by pollen and spores in the Beardmore Glacier region as Late Triassic (Farabee, Taylor, and Taylor 1989).

One coal seam was studied in detail over a distance of 10 meters. The studied sequence consists of a basal silty layer, a thin carbonaceous underclay overlain by the coal, and an uppermost fine- to medium-grained sandstone with shaly lenses (figure 1). The coal seam shows a distinct hummocky topography (figure 2). These hummocks are localized accumulations of permineralized, thin (3–6-centimeter) compressed logs alternating with coalified horizons. None of the logs was found in growth position, and no distinct rooting horizon was observed. Depressions have few permineralized logs or woody debris and are mainly coal.

Because highly decomposed peat compacts more than dense, woody particles do, differential compaction after burial might have enhanced the observed relief, but it cannot completely account for the ridge-and-swale structures. The overlying sandstones thicken in depressions and thin above the mounds, indicating that this microrelief predated deposition of the cover sequence. Fine, organic-rich clayey layers are restricted to areas above depressions, as if fine sediment settled preferably in lower lying areas. Erosion can also be discounted as a relief-former because individual peat layers can be traced laterally without interruption, and cut-and-fill structures are not associated with development of the ridges and swales. Absence of sand or gravel in association with logs and woody debris rule out transport and deposition within a fluvial system. The process responsible for the recurring scheme of alternating logs and peat layers within distinct hummocks must have taken place in a relatively stable environment because time for formation for these deposits was at least 6,000 to 10,000 years, using normal rates of peat accumulation (Retallack 1990). Therefore, distinct ecological conditions were the primary relief-forming mechanism for development of these structures.

Today, forested fens or bogs constitute the margins of minerotrophic mires (Tallis 1983). Mire-fringes are often invaded by trees that form on elevated, drier tops of tussocks (Tallis 1983). As the weight of the tree increases, the tussocks become unstable and eventually sink or roll over, killing the

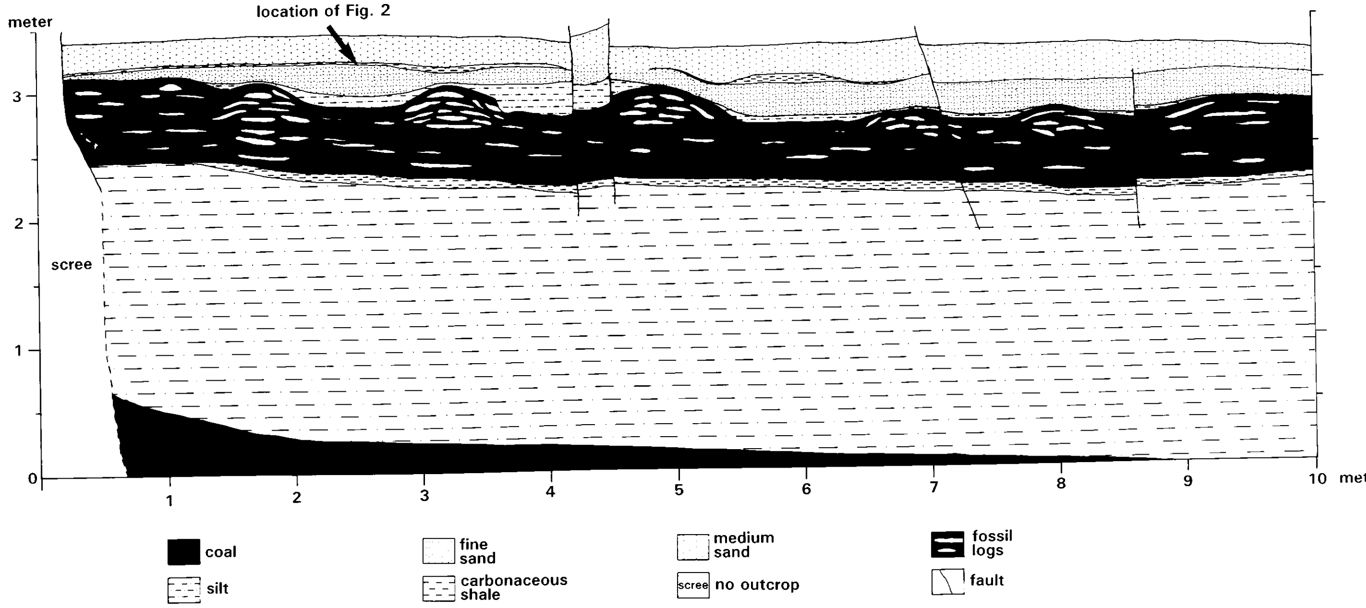

Figure 1. Lateral section of coal seam and enclosing strata, 10 kilometers southeast of Schroeder Hill, central Transantarctic Mountains, Antarctica.

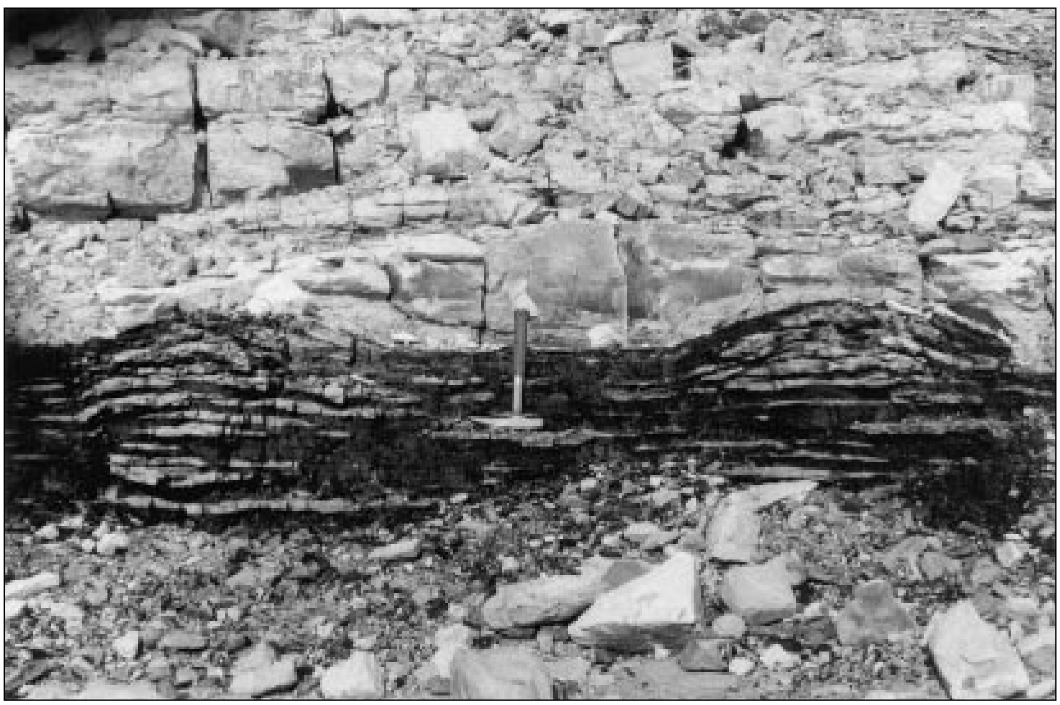

Figure 2. Hummocky topography of a coal seam near Schroeder Hill, central Transantarctic Mountains, showing domes on left-hand side of figure 1.

trees. This zone is also exposed to high winds. Young, shallow-rooted trees at the mire margin are more susceptible to windthrow and are easily uprooted before maturing (Allen 1992; Ehrenfeld 1995). Thus, this zone is characterized by many young dead or dying trees on depressed waterlogged tussocks. These logs from windthrow or drowning provide suitable regeneration sites because of the elevated microtopography (Ehrenfeld 1995) and the increased nutrient availability (Agnew, Wilson, and Sykes 1993). Growth of mosses on the fallen tree in the primarily waterlogged environment forms a new peat-layer. At these sites of greater nutrient availability and drier conditions, vegetation growth is favored (Ehrenfeld 1995). As the initial microrelief increases, floras less-adapted to waterlogging colonize the site as the hummock rises above the water table. Continuous peat accumulation on fallen logs, preferred growth of young trees on elevated sites, and episodic windthrow and drowning lead to the formation of hummocks with alternating peat and wood layers.

Today, similar ecotones of forested peatland and mires outside the zone of continuous or discontinuous permafrost occur at latitudes of 40° to 55° in humid, cool, temperate climate zones. During the late Triassic, Schroeder Hill was situated at comparable latitudes of 57° to 55°S (Scotese 1994), indicating that paleoclimate belts in the midlatitudes of the late Triassic were similar to today's climate zones.

I thank Shaun Norman for assistance in the field and Greg Retallack for helpful comments. Especially appreciated were the skilled pilots of Helicopters New Zealand and the organizational efforts of David Elliot and Kevin Killilea in the Shackleton base camp of 1995–1996. This work was supported by National Science Foundation grant OPP 93-15228.

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Geologic studies on rocks of the Jurassic Ferrar Group

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As part of the helicopter-supported research in the Shackleton Glacier region, geologic studies were conducted on Jurassic extrusive and intrusive rocks (Ferrar Group) and, to a limited extent, on underlying Permian and Triassic strata (Beacon Supergroup).

Work on the Ferrar rocks had two principal aims: first, the collection of Ferrar Dolerite sills for geochemical analysis and radiometric dating and, second, the examination and collection of rocks from the Prebble Formation, a pyroclastic sequence underlying the Jurassic Kirkpatrick Basalt, in order to understand its paleovolcanology and tectonic setting. Dolerite sills were collected at sites from Cape Surprise to Otway Massif (figure) to provide as complete a stratigraphic coverage of their occurrence as possible. Sites ranged from the basement granite to the Prebble Formation, thus encompassing the whole of the Beacon sequence. In addition, a special effort was made to collect sills from the Nilsen Plateau that have been reported to carry significant amounts of olivine and, thus, possibly to be more mafic than any other known occurrence of Ferrar tholeiites. In the course of work on the Otway Massif, a subvolcanic system was discovered. This system shows subsurface interactions related to the formation of the pyroclastic rocks of the Prebble Formation.

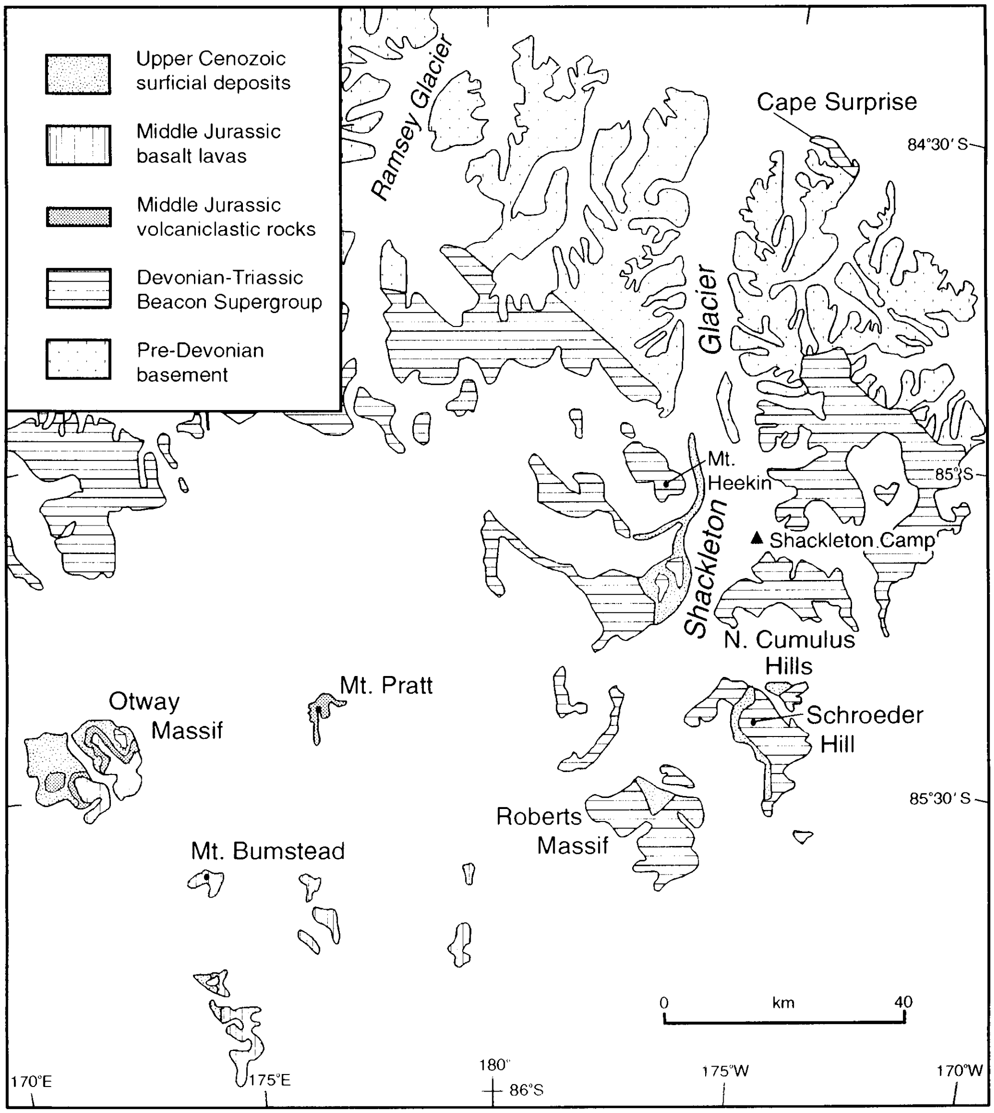
The pyroclastic rocks of the Prebble Formation were examined at Mount Pratt and Otway Massif where as much as 360 meters (m) of stratigraphic section are present (Elliot and Larsen 1993). New sections were measured, and previously measured sections were re-examined; evidence was found for water-magma interactions, and these interactions support a phreatomagmatic origin for these deposits, like that advocated for the Prebble Formation rocks in the Marshall Mountains north of the Beardmore Glacier (Hanson and Elliot 1994, in press). Based on regional relationships, the contact with the Triassic Falla Formation appears to have significant topographic relief. The Falla Formation shows considerable tilting and folding, both of which are attributed to the emplacement of Ferrar magmas at shallow depths. Intrusion breccias with compositions somewhat similar to the breccias of the Prebble Formation occur in a number of places as thin dikes and small bodies. At two localities, the contact with the overlying basalts is marked by a paleosol. In the course of work on the Beacon rocks, an additional area of deformed Falla Formation strata was discovered near Schroeder Hill and again is attributed to emplacement of Ferrar magmas at shallow depth; although intrusion breccias were not found, dikes with irregular and contorted forms cut the Falla rocks.

Coarse-grained channel sandstones were collected from the Upper Permian part of the Buckley Formation and from the Triassic Fremouw and Falla Formations for provenance studies on zircons and thermal history from apatites. The former will be used to investigate evidence for an active magmatic arc as well as crustal ages in the source regions and the latter, for estimates of the possible original stratigraphic thickness of the lavas as well as timing of any postdepositional heating.

Field checking of the geology of the Shackleton and Ramsey Glaciers region was carried out so that, with the aid of existing maps and information supplied by other project participants at the Shackleton Glacier camp, three 1:250,000 geologic quadrangle maps can be completed.

After completion of the work in the Shackleton Glacier region, further fieldwork was conducted at the Coombs Hills in southern Victoria Land where pyroclastic rocks of the Mawson Formation, equivalent to the Prebble Formation, are extensively exposed (Grapes, Reid, and McPherson 1974; Bradshaw 1987). The Mawson rocks show similarities to the Prebble Formation breccias and again are the products of major phreatomagmatic eruptive events that involved explosive interaction between magma and groundwaters held in the Beacon sedimentary rocks. Clastic dikes, made up of varying proportions of sedimentary and volcanic material, are common, as are intrusion breccias and small plugs and dikes of basalt. The great thicknesses of breccia and the widespread occurrence of dikes and sills in the Mawson Formation suggest that the region was a major eruptive center during the Jurassic.

Particular thanks are given for the logistic support provided by U.S. Navy VXE-6 squadron and Helicopters New Zealand. David Buchanan and Robert Andress provided invaluable assistance in the field. This field research was supported by National Science Foundation grant number OPP 94-20498 to Ohio State University.



Location and simplified geologic map of the Shackleton Glacier region. Ferrar Dolerite sills are co-extensive with both the Permian-Triassic Beacon Supergroup rocks and the Middle Jurassic volcaniclastic rocks of the Prebble Formation. Solid triangle indicates the location of the Shackleton camp.

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Three-dimensional magmatic filling of Basement sill revealed by unusual crystal concentrations

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When magma moves from great depth to the upper reaches of Earth, the vigor and nature of its motion can sometimes be traced through the entrainment and distribution of unusually large crystals. Called *phenocrysts*, these crystals and their distribution are nowhere better exhibited than in the Basement sill, a thick [350-meter (m)] layer of magmatic rock, exposed throughout the McMurdo Dry Valleys region. What is even more noteworthy is the great abundance and three-dimensional distribution of these phenocrysts. This distribution promises to reveal both the initial magmatic injection point as well as the lateral flow field over an area of thousands of square kilometers. Discerning and understanding this flow field, an unprecedented achievement, give fundamental insight into the essential link between volcanism and magmatism and the dynamics of magma beneath ocean ridges the world over. After giving a brief background on the nature of crystal entrainment and sorting in ascending magma, we present the results of detailed sampling and chemical analysis from seven sections of the Basement sill covering much of the McMurdo Dry Valleys.

Heavy solid particles in fluids coursing through pipes and slots tend to migrate away from the walls and form high concentration sluglike flows in channel centers. The behavior of fatty globules in blood and paper pulp paste particles in aqueous solutions are common examples of this process. Similarly, magma ascending sufficiently fast through packed beds entrains crystals, if it moves fast enough, carrying them upward. These crystals are almost always heavy, and they migrate away from the walls to the flow center, but they also sink *en mass*, relative to the rising magma. The net result is a rising column of magma led by essentially crystal-free magma and followed by a crystal-rich tongue of increasingly coarse crystals. The flow sorts the crystals vertically and laterally such that the distribution of crystals reveals the flow field; the longer the magma carries the crystals, the better the sorting (*see* figure 1).

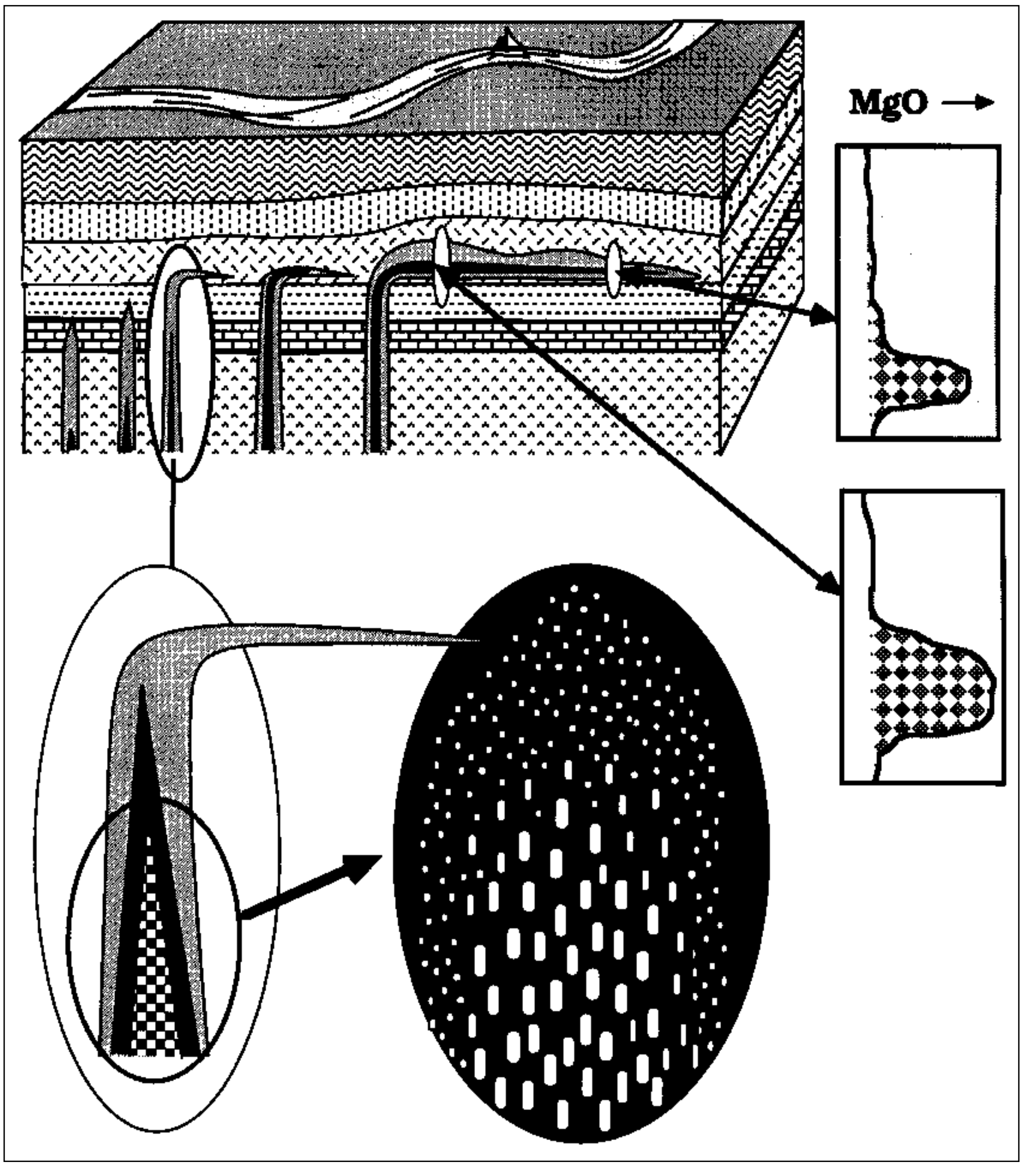

Figure 1. A schematic depiction of the process of crystal entrainment and sorting in an ascending body of magma. The sequential process of emplacement of the magma as a horizontal sheet or sill of magma is also shown along with possible variations in magnesium concentration due to sorting of the entrained crystals during flow.

This phenomenon is well known in fluid mechanics (e.g., Segre and Silberberg 1962; Lael 1980) and has been nicely modeled as a magmatic process by Bhattacharji and Smith (1964), who have also noted its importance in forming the Muskox intrusion in northern Canada. Simkin (1967, pp. 64–69) and Upton and Wadsworth (1967) have shown the prevalence of this style of sorting or flow differentiation in, respectively, dikes in Scotland and a small sill on Reunion Island. Gibb (1976; *see also* Marsh 1996) has drawn attention to the possibility that this kind of magma may go to form large, well-sorted magmatic bodies that have been routinely attributed to *in situ* crystallization, but until now direct evidence has been lacking. The critical evidence is preserved over great expanses of the Basement sill.

The presence of unusual concentrations of large crystals of orthopyroxene (opx) in the Basement sill was first reported by Gunn (e.g., 1966). He noted the curious complete absence of opx along the upper and lower margins even when a relatively thick (30-m) layer of large opx crystals forms the sill center at Solitary Rocks. Realizing the possible opportunity presented by this opx, we have begun mapping out this concentration in three dimensions by studying and collecting samples at (so far) eight locations. The vertical variation of magnesium oxide (MgO) (weight percent, wt.%), which is a direct reflection of opx content, through the Basement sill at seven of these locations is shown by figure 2.

Over a distance of about 15 kilometers (km) in the northeast wall of Wright Valley, the sill pinches out from a thickness of over 350 m near Bull Pass to a thickness of 5 centimeters (cm) at the north lower Wright Valley section. This captures the leading edge of the invading magma, and this easternmost section contains no sign of opx; MgO is constant at the background level of about 7 percent. Several kilometers west, approaching Bull Pass, the concentration steadily increases to over 10 percent MgO in the sill center (see Mount Peleus section), to over 15 percent at east Bull Pass, and to 20 percent MgO at west Bull Pass. The opx tongue fills about 75–80 percent of the sill and individual crystals reach lengths of 8–10 millimeters (mm), whereas at the leading eastern edge the tongue is thin and opx size is 1–3 mm. Twelve kilometers farther east from Bull Pass in the Dais section, although incomplete due to debris cover, the sill resembles the west Bull Pass section if allowance is made for a confused zone at the upper contact. Ten kilometers north (Victoria Valley section) and 30 km south (northwest Kukri Hills section) from west Bull Pass the opx tongue is still prominent, although it is, especially in the Kukri Hills, less strong (*see* figure 2).

These data suggest that somewhere, perhaps near present Lake Vanda, the Basement sill began filling and magma spread laterally in all directions while progressively lifting the overlying 4 km of crust. The vigor of emplacement was strong enough to entrain a considerable mass of large orthopyroxene crystals whose distribution intimately records this event. Although it is still premature to draw more conclusions, judging from the internal structure of the tongue, filling may have been pulsative as in volcanism, and filling may have been stronger along than transverse to the Transantarctic Mountains (i.e., north-south), which is similar to the style of magmatism at ocean ridges.

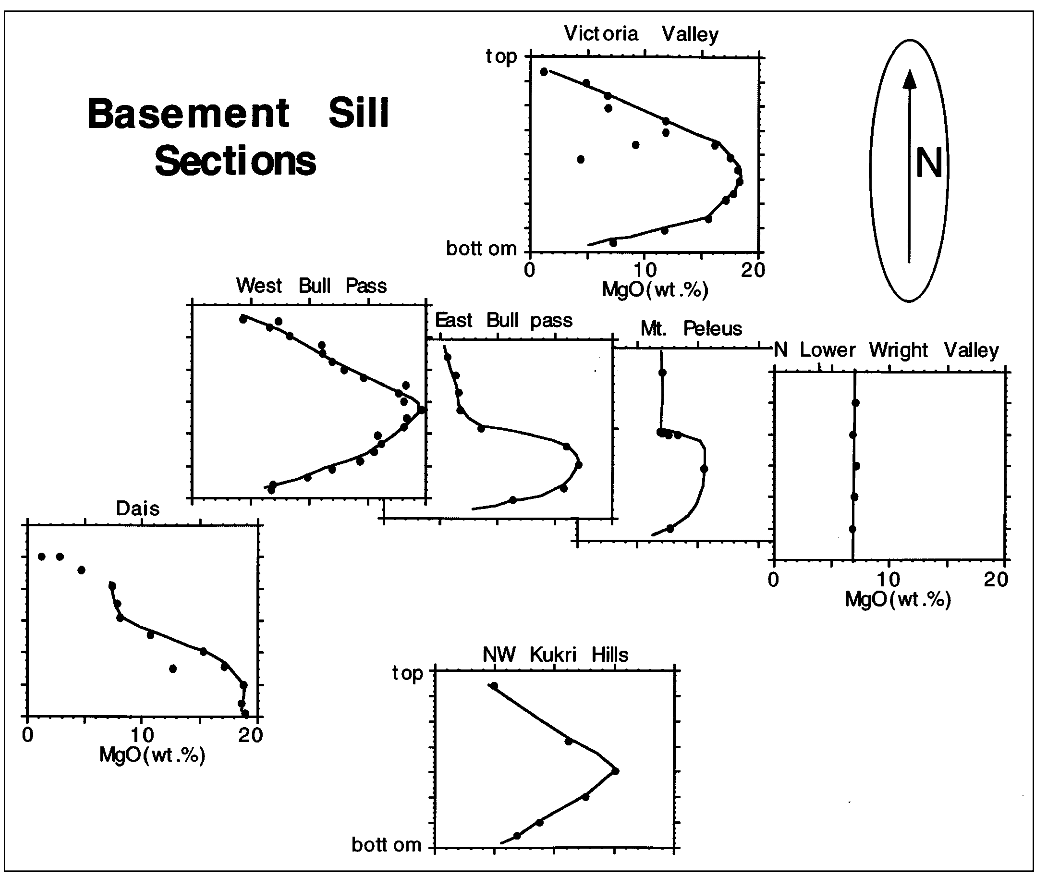

Figure 2. The variation of MgO (wt. %) concentration through seven vertical sections of the Basement sill in the McMurdo Dry Valleys region. The location of each section is as labeled and the series of sections is depicted in relative geographic position north-south, east-west. The few points in the Victoria Valley section falling well behind the curve reflect merely local concentrations of other minerals, mostly plagioclase, due to internal sorting within the opx tongue.

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Digital recording of the seismicity of Mount Erebus volcano, November 1994 to June 1996

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The Mount Erebus Volcano Observatory project was established in 1992 to provide digital monitoring and recording of seismic activity associated with the volcanism of Mount Erebus. The volcano is the central feature of Ross Island, with the summit crater located approximately 35 kilometers (km) from McMurdo Station. A permanent convecting crater lake of anorthoclase phonolite magma produces frequent small strombolian eruptions (Kyle 1994). Data from the radiotelemetered seismic network (figure 1) are digitized and recorded at 100 samples per second using an event-triggered PC-based detection and recording system at McMurdo Station. Data are compressed and automatically transferred daily over the Internet to New Mexico Institute of Mining and Technology and Victoria University for analysis (Skov, Kyle, and Aster 1994). During the study period (November 1994 through June 1996), over 3,000 digitally recorded events were recorded and examined. Continuous analog helicorders are also operated at McMurdo, and a count of events recorded during 1995 was used to examine digital data acquisition efficiency (figure 2). Analog records indicate that, although many seismic events currently occurring on Erebus are of insufficient size to trigger the digital acquisition system, the larger events which trigger the system do provide a representative sample of the frequency and style of the overall seismic activity. At least four categories of seismic events are recognized: long-period (LP) events, volcano-tectonic (VT) events (Chouet 1996), explosive (E) events (Dibble, O'Brien, and Rowe 1994), and long-period tremor (Chouet 1985). The system also records near-regional, regional, and teleseismic signals originating far outside of the array.

Locations

Preliminary locations for the three principal seismic event types (figure 3) were obtained using a gradient-over-a-half-space velocity model estimated from refraction data by

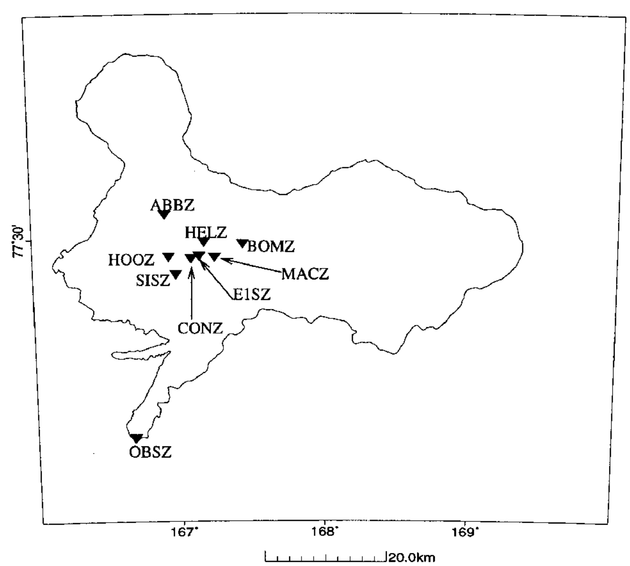

Figure 1. Seismic stations in the current Mount Erebus Seismic Network. Eight of these stations are located at elevations ranging from 1,789 meters to 3,708 meters and one (OBSZ) is installed near McMurdo Station at an elevation of 30 m. SISZ was upgraded to three components during the 1995–1996 field season; all other stations have vertical-component, 1-Hz seismometers.

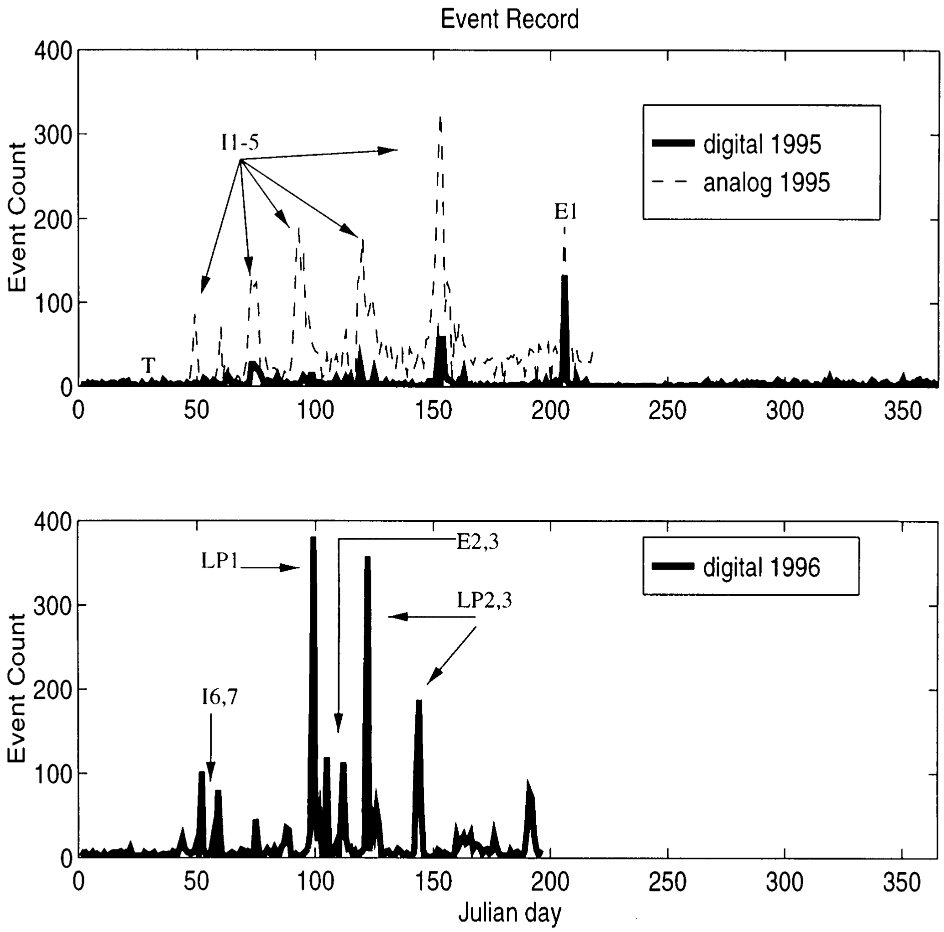

Figure 2. Daily histogram of recorded events for 1995 and 1996, including both analog and digital counts for part of 1995. T indicates an approximately 15-minute-long period of high-amplitude, continuous tremor on 31 January 1995. I1-5 indicates 5 successively bigger swarms of shallow earthquakes (possible ice-quakes) which occurred during austral autumn of 1995. E1 indicates a swarm of 150 digitally recorded summit Strombolian explosion events. Note that there is some bias in counting statistics imposed by the digital system, which is more sensitive to swarms containing larger events that are more likely to carry the network (e.g., I5 compared to E1). I6,7 indicate two shallow swarms in early 1996, and LP1-3 mark swarms of both shallow and LP events. E2,3 indicate two swarms of explosion events which occurred one week apart in April of 1996.

Rowe (1988). The events plotted met the following requirements.

- Seven or more stations clearly recorded the event.
- First arrivals are sufficiently impulsive to estimate phase arrival times with a 1-sigma uncertainty of less than 0.2 seconds.
- The hypocenter solution is well-modeled (a root mean square residual value not greater than 0.2 seconds).
- The hypocenter estimate is within approximately twice the network aperture and within the volcano.

VT events

VT events are indicative of the faulting of brittle rock due to perturbations caused by the intrusion or withdrawal of fluids superimposed on the regional stress field. Because they originate in abrupt shear motion along faults, VT events exhibit sharp arrivals for both compressional-wave and shear-wave phases at multiple stations. Only 87 out of the more than 3,000 events examined show clear VT characteristics. We estimate the relative sizes of seismic sources using a simple scale based on the duration of the seismic signal. The mean duration magnitude of located VT events is 1.04 and the majority occur 1 km or more below the summit (figure 3, *top*). VT events do not occur in temporal swarms, arguing for steady-state stress conditions in the lower volcano.

LP events

LP events are related to magmatic source processes and are characterized by emergent first arrivals, relatively long-period sources [typically below approximately 2 hertz (Hz)], and a general deficit of identifiable shear-wave arrivals. LP events make up the largest percentage of the total data set (approximately 65 percent). Over 500 LP events occurred during seismic swarms, accounting for more than one-quarter of the total number of LP events. LP events are commonly difficult to locate accurately due to highly emergent onsets. Additional complications arise in locating events from LP swarms due to the interleaving of arrivals from multiple events, and swarm location estimates are thus not presently included in the location catalog. LP event locations are scattered and are predominantly distributed throughout the upper 1 km of the mountain (figure 3, middle), suggesting that the upper mountain is criss-crossed with numerous small pathways and conduits for movement of magma. The mean duration magnitude of located LP events is 1.97.

E events

E events are directly attributable to volcanic explosions at or beneath the surface of the lava lake. Low-frequency precursors frequently accompany E events and may be caused by oscillations of exsolved gas bubbles prior to the

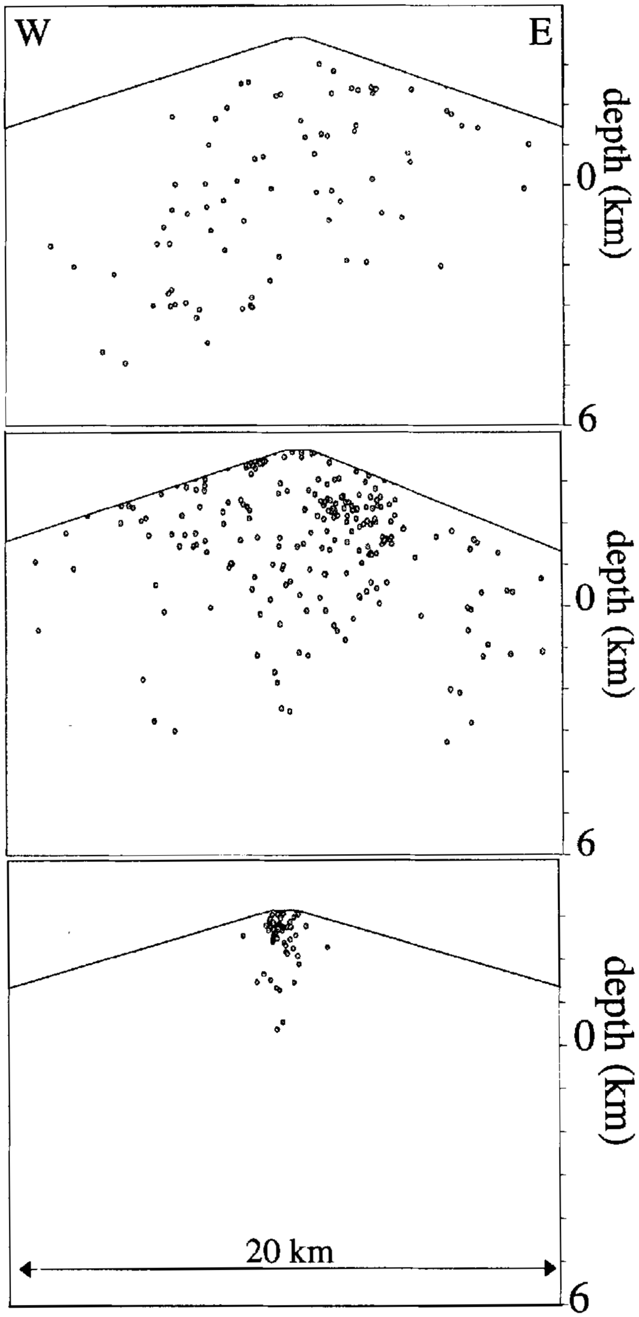

Figure 3. West-to-east cross sectional views of VT- (*top*), LP- (*middle*), and E-type (*bottom*) events.

explosion (Dibble et al. 1994, pp. 1–16). E events constitute approximately 20 percent of the total data set. Like LP events, E events are observed to form temporal swarms; over 230 E events occurred during two main periods of explosion swarms, accounting for over one-third of all such activity. E sources locate in the summit lava lake area and in the upper portion of the main conduit (figure 3, bottom). The mean duration magnitude of E events for which a coda could be picked is 0.74. This number is unrepresentative of the population of larger explosion events, however, because the current trigger parameters terminate many of the digital records prior to the time that the coda fades into the background.

Two especially notable swarms of E events occurred during the study period. On 14 April 1996 (julian day 105; figure 2) a swarm of 100 triggered E events occurred near the active crater. A second swarm of 165 triggered E events began on 20 April 1996 (julian day 111; figure 2). Earthquake occurrence rate versus magnitude statistics for these two swarms indicates that a relatively large number of small events occur during E swarms compared to most types of seismic activity, a feature previously noted by Dibble et al. (1984).

Long-period tremor

An episode of long-period harmonic tremor was recorded on 30 January 1995. The dominant frequency of the tremor changed during the 5 minutes that the tremor was recorded from approximately 5 Hz at the start of the record to approximately 1.5–2 Hz after several minutes. This change could be due to a widening conduit or to changing flow conditions. The dominant interpretation of such tremor episodes (e.g., Chouet 1985) is that they result from the superposed resonant responses of a volcanic conduit to numerous individual pressure pulses during periods of magmatic transport. Thus, it is possible that this episode represents a significant upwelling of magma from depth. The observation that the tremor was clearly recorded even at the McMurdo seismic station, located 35 km from the summit, indicates a deep focus and large-amplitude source.

Acknowledgments

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Global positioning system static sites along the Victoria Land coast of Antarctica

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The Victoria Land coast of Antarctica has raised beaches with fossil deposits that reflect eustatic sea-level and west-antarctic-ice-sheet changes since the last glacial maximum. During the 1994–1995 austral summer, these raised marine features were investigated from McMurdo Sound (approximately 78°S) to Terra Nova Bay (approximately 74°S) by

- collecting fossils for subsequent geochemical analysis,
- analyzing the stratigraphy of the fossil deposits,
- assessing the surrounding glacial geomorphology, and
- surveying the fossil locations with Trimble 4000 SSE dual-frequency global positioning system (GPS) receivers.

Permanent GPS site markers were installed by embedding "Bevis" pins into bedrock to create static sites for local kinematic surveys along the Victoria Land coast (figure). Dual-frequency phase data from these static sites were collected for 8-hour periods with 15-second epochs and post-processed in the World Geodetic System 1984 (WGS84) reference frame. An orbit determination model produced by the Department of Geodetic Sciences at Ohio State University, GPS Orbit Determination Including Various Adjustments (GODIVA), was used for calculating the static-site positions (table). These static-site position determinations had the same level of precision (standard deviations in the table) as has been reported for the continuously recorded International GPS Service for Geodynamics (IGS) site at McMurdo Station, which ran through 1994 (Hothem personal communication).

Geodectic comparisons between the static sites at McMurdo Station and Explorers Cove, which was anchored into the cement casing surrounding the drill-core stem from holes 8–10 of the Dry Valley Drilling Project (McKelvey 1985, pp. 63–94), were made to assess the reliability of orthometric heights which were calculated from modeled geoid heights and measured ellipsoid heights (Heiskanen and Moritz 1987; equation).

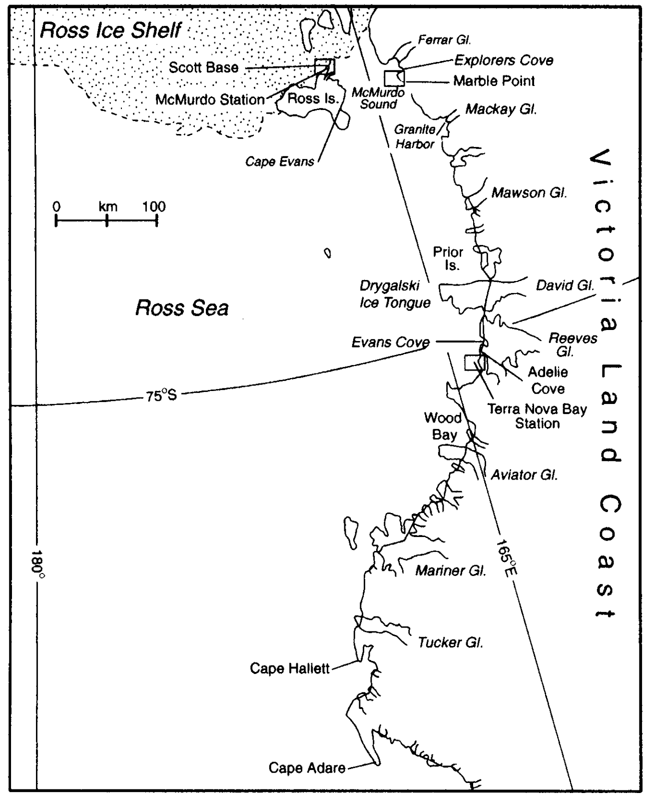
$$(h_1 - h_2) - (N_1 - N_2) = (H_1 - H_2) \tag{1}$$

where

h is the height above the reference ellipsoid as determined from the GPS surveys;

H is the calculated orthometric height above the geoid; and

N is the modeled geoid undulation for sites 1 and 2 derived from OSU91a (Rapp 1992).



Map of the Victoria Land coast in the Ross Sea, Antarctica, showing the general locations of the GPS static sites that are referenced in the table.

The agreement between the measured and calculated orthometric heights (table) provides support for the geoid model (OSU91a) in the Ross Sea region of Antarctica.

The static sites in the table were used for kinematic surveys of the positions and elevations of more than 100 marine-fossil sites that were identified in Explorers Cove and Marble Point, raised-beach strandlines and fossiliferous marine terraces in Marble Point and other coastal regions, and various glacial geomorphological features from McMurdo Sound to Terra Nova Bay. In the future, these static sites along the Victoria Land coast may prove useful for interpreting crustal motion responses to changes in the west antarctic ice sheet after the last glacial maximum (Elliot, Strange, and Whillans 1991; James and Ivins 1995).

We would like to thank Brett Baker, Olafur Ingólfsson, Skip Van Bloem, Mike Prentice, and Kazuomi Hirakawa for their assistance in the field. This research was supported by a grant from the National Science Foundation (OPP 92-21784) to P.A. Berkman.

Static site coordinates along the Victoria Land coast, Antarctica[a]

Station name	Latitude south (DMS)[b]	Longitude east (DMS)[b]	Measured ellipsoid height (m)	Modeled geoid height (m)[c]	Calculated orthometric height (m)	Measured orthometric height (m)[d]	Survey days
McMurdo Station (McMu)	77°50'55.127" ±0.3493E-03"	166°40'31.135" ±0.1555E-02"	–1.0553±0.0249	–54.581	53.526	53.22[f]	6
Explorers Cove Rock (VCE943)	77°33'21.083" ±0.6461E-03"	163°31'18.337" ±0.2085E-03"	–39.9256±0.0377	–55.217	15.291	—	2
Explorers Cove drill stem (VCE944)	77°34'38.448" ±0.2271E-03"	163°31'5.530" ±0.6525E-03"	–51.7300±0.0306	–55.193	3.463	2.8[g]	4
Marble Point (WALT)	77°25'56.947" ±0"	163°49'15.082" ±0"	–22.4758±0	–55.339	32.863	—	1
South Stream (VCE946)	77°27'3.237" ±0.8258E-03"	163°44'46.385" ±0.3160E-02"	–42.6855±0.0722	–55.320	12.635	—	3
Gneiss Point (VCE948)	77°25'11.295" ±0.6459E-03"	163°44'12.846" ±0.2323E-02"	–46.4402±0.0226	–55.375	8.935	—	2
Evans Cove (VCE949)	74°52'45.349" ±0.3028E-03"	163°55'29.889" ±0.2741E-02"	–34.2381±0.0724	–60.682	26.444	—	4
Adélie Cove (VCE9410)	74°45'19.275" ±0.2050E-03"	163°59'23.540" ±0.1330E-02"	263.1833±0.0172	–60.323	323.506	—	3
Terra Nova Bay Station (NOVA)	74°41'38.949" ±0.4423E-03"	164°6'49.758" ±0.1286E-02"	–41.0303±0.0426	–60.193	19.163	—	3

[a]The GPS dual-frequency phase data and the metadata for the above static sites are archived at the University NAVSTAR Consortium (UNAVCO) facility, which is supported by the National Science Foundation, in Colorado.
[b]DMS denotes arc degrees, minutes, and seconds.
[c]Based on geoid undulation model OSU91a (Rapp 1992).
[d]Measurements available only for McMurdo Station and Explorers Cove static sites.
[e]Minimum 8-hour continuous record each day.
[f]Measurement from Hothem (personal communication).
[g]Measurement from McKelvey (1985, pp. 63–94).

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Re-evaluation of the structure and stratigraphy of the Heritage Range, Ellsworth Mountains

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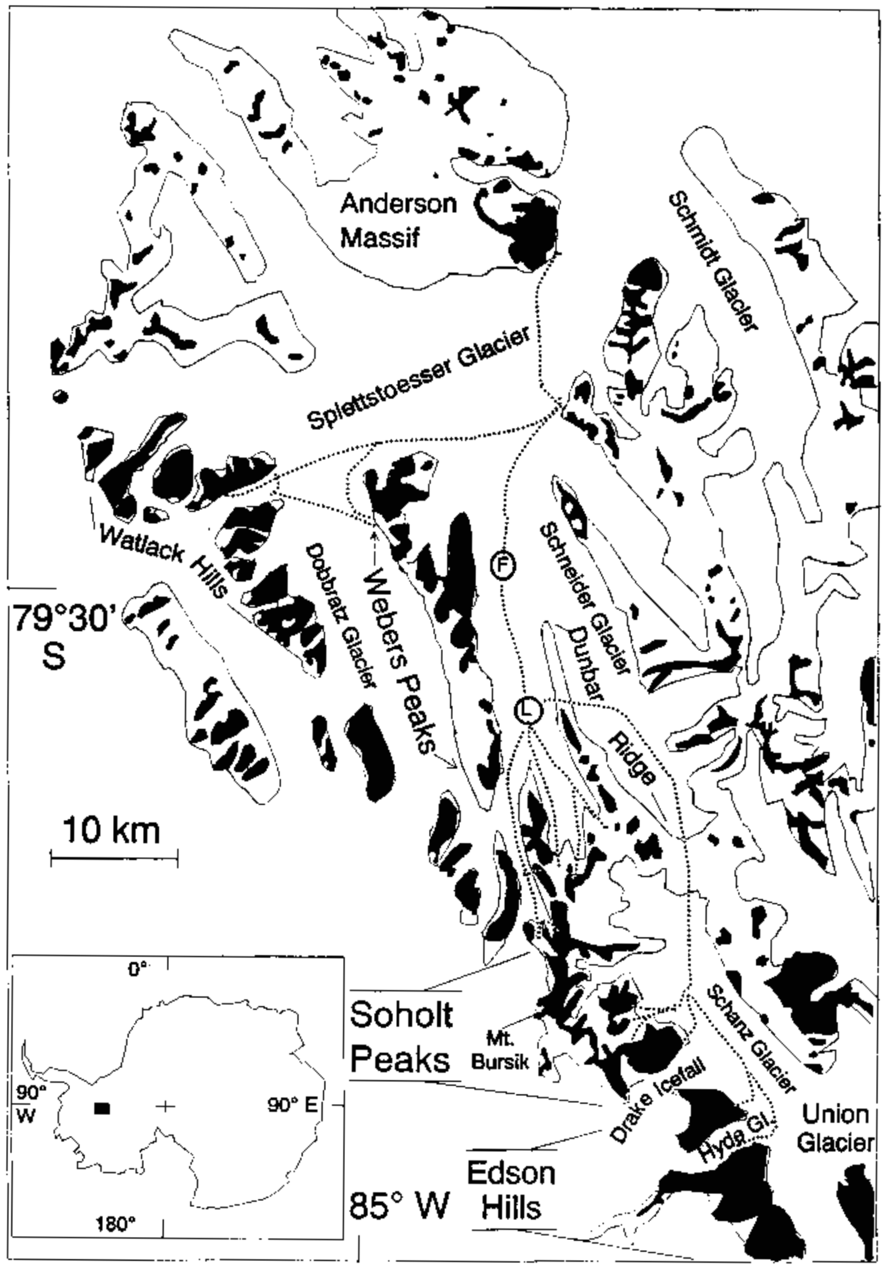
The Ellsworth Mountains are part of a geologically and geophysically defined terrane that lies between the Transantarctic Mountains and West Antarctica. The approximately 13,000-meter-thick succession of Cambrian through Permian strata exposed in the Heritage Range has paleogeographic and paleobiogeographic affinities to the margin of Gondwanaland (Webers, Craddock, and Splettstoesser 1992), but the Early Paleozoic depositional and tectonic history of the terrane is equivocal (Thorstenson, Duebendorfer, and Rees 1994; Curtis 1995; Rees et al. 1995). Establishing the geological history of the range is critical for assessing plate tectonic models of the paleo-Pacific-facing margin of Gondwanaland.

Our fieldwork during the 1995–1996 season was concentrated in the northern Heritage Range of the Ellsworth Mountains (figure). Our field party consisted of Ernest M. Duebendorfer (structural geologist), Margaret N. Rees (stratigrapher), Eugene Smith (volcanologist), and Lucylle J. Smith (mountaineer). On 24 November 1995, an LC-130 aircraft with a VXE-6 crew airdropped four 55-gallon drums of motor-gas on the Balish Glacier during our reconnaissance flight, and on 29 November, they put our party into the field near the drop (79°32'14"S 84°25'11"W). Using four snowmobiles and four Nansen sledges, we traversed the area and established temporary camps. Our party, fuel drums, gear, and rock samples were pulled out of the field on 3 January 1996.

Our geological mapping documents at least three deformational events in the northern Heritage Range, not just the single Mesozoic contractional event recognized by other workers. Furthermore, our recognition of structures not previously mapped calls into question the stratigraphic order, ages, and depositional relationships of several map units.

The earliest documented deformational event (D_1) is manifested by refolded folds and crosscutting cleavages in the Middle Cambrian Springer Peak Formation. D_1 structures are nearly completely overprinted by the dominant fabric of the range. A minimum age for D_1 may be constrained by the angular unconformity between the Late Cambrian Minaret Formation and the overlying Crashsite Group in the Webers Peaks area. The contact



Generalized map of the northern Heritage Range, Ellsworth Mountains, Antarctica. Solid black denotes exposed rock; black outline denotes glacier-rock interface; F denotes fuel depot; L denotes landing site; dotted line denotes path of field party.

previously was described as conformable by Webers et al. (1992) and disconformable by Goldstrand et al. (1994). The unconformity is bracketed between post-early Late Cambrian and pre-Devonian based on fauna that we collected immediately above and below the contact (Rees et al. 1995). Thus, deformation of Cambrian succession may predate or have been coeval with the development of this unconformity. The D_1 event and the unconformity may be related to either the Ross Orogeny, activity along the Mozambique suture, or other movements along the paleo-Pacific margin of Gondwanaland.

The dominant structures in the Heritage Range are north-northwest-striking folds associated with the Triassic Ellsworth/Gondwanide Orogeny (D_2). Our work and that of Curtis (1995) suggest that structures attributed to this event record not only northeast-southwest shortening but also involve a significant component of dextral shear deformation.

In the Soholt Peak–Edson Hills region of the northern Heritage Range (figure), we documented a series of east-vergent thrust sheets associated with D_2. Several contacts between formations that were originally interpreted as depositional are major thrust-sense, cataclastic shear zones that dip 40–60° west. Kinematic indicators record dextral, top-to-the-east tectonic transport. The structurally lowest of these thrust sheets places the lower Middle Cambrian Drake Icefall Formation over the undated Union Glacier and Hyde Glacier Formations along the Drake Icefall shear zone. The structurally intermediate-level Conglomerate Ridge shear zone places the Conglomerate Ridge Formation in the hanging wall over the Drake Icefall thrust plate. The structurally highest shear zone places rocks originally mapped as Springer Peak Formation against the Conglomerate Ridge Formation.

An additional structure in the Soholt Peaks area is an inferred cross fault that strikes northeast from Mount Bursik (figure). This structure is inferred on the basis of right-lateral separation of formation contacts and of right-lateral displacement of a regional anticline. If this interpretation is correct, then the cross fault has a minimum of 5 kilometers of right-slip displacement.

A third, possibly late- or post-Gondwanide deformational event is suggested by the orientation of F_2 folds. F_2 fold hinges show a strong bimodal distribution suggesting that they have been refolded about subhorizontal, northeast-trending fold axes (F_3). In addition, a set of northeast-striking, subvertical joints is developed throughout the northern Heritage Range. The mean orientation for these joints is similar to the orientations of the inferred F_3 fold axes. This joint set may therefore represent axial planar tension joints associated with the northeast-trending F_3 fold axes.

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